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Coupled hydrological and biogeochemical dynamics in high elevation meadows: thresholds, resiliency and change

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COUPLED HYDROLOGICAL AND BIOGEOCHEMICAL DYNAMICS IN HIGH ELEVATION MEADOWS: THRESHOLDS, RESILIENCY AND CHANGE

A dissertation submitted in partial satisfaction of the requirements for the degree Doctor of Philosophy

in

Environmental Systems

(Soil Biogeochemistry and Soil Physics)

by

Chelsea Lynn Arnold

Committee in charge:

Asmeret Asefaw Berhe - Co-Chair

Teamrat A. Ghezzehei - Co-Chair

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2014

The dissertation of Chelsea Lynn Arnold is approved, and is acceptable in quality and form for publication on microfilm and electronically:

Asmeret Asefaw Berhe

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Eric L. Berlow

Samuel J. Traina

University of California, Merced

2014

DEDICATION

To the amazing women in my life who gave me strength and encouraged my dreams…

my mother, Gail Talmadge,

my sister and best friend, Keaton Koechli,

my aunt, Jo Nunes

my grandmother, Barbara Talmadge,

my grandmother, Clara Nunes,

my great-grandmother, Huldah Tichenor

my mother in law, Sallie Arnold

my step mother, Rosie Talmadge

and

my daughter, Mackenzie Arnold

To my husband and son who fill me with love

Buz Arnold and Owen Arnold

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Acknowledgements

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I would like to thank my husband, Buz, for his endless support and understanding as I gave up a teaching career to pursue this dream of getting my Ph.D. My children, Mackenzie and Owen, you grew up watching your mother get a second chance at her dreams and I can only hope it will give you inspiration to pursue yours, whatever they may be.

My dissertation was funded in part by the startup funds of Asmeret Asefaw Berhe and Teamrat A. Ghezzehei, the Mildred E. Mathias grant, and the Hellman Family fund and the University of California, Merced; for all of which I am eternally grateful.

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PLOS One: Submitted

A method for characterizing desiccation induced consolidation and permeability loss of soft soil 2014
In preparation

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Nature Education, in press

Exceedance of historic limit of dryness triggers hydrological tipping point in high elevation peatlands 2014
In preparation

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Impact of lowered water table on meadow water holding capacity
Mathias Research Symposium 2012
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Changes in soil structure in response to extreme drought in the High elevation meadows of Yosemite National Park
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Abstract of the Dissertation

Coupled Hydrological and Biogeochemical Dynamics in High Elevation Meadows: Resiliency, Thresholds and Change

by

Chelsea Lynn Arnold

Doctor of Philosophy in Environmental Systems

(Soil Physics and Biogeochemistry)

University of California, Merced, 2014

Asmeret Asefaw Berhe – Co-Chair

Teamrat A. Ghezzehei – Co-Chair

High elevation meadow ecosystems play a fundamental role in the storage and movement of water from the snowpack to the streams. The vulnerability of such systems to extreme changes in the depth and duration of snowpack in a given year is unknown. However, due to the coupled nature of the hydrological and biogeochemical cycles in the meadows, it is expected that a change in hydrology will impact biogeochemistry and vice versa. This research investigates the gaps in our knowledge of the coupled hydrology and carbon cycling of meadow soils, and forges new territory in our understanding of the impacts of lowered water tables and desiccation on the hydrologic resiliency of those soils. This knowledge is critical in snowmelt-dominated watersheds, where meadows (both high elevation and montane) serve as natural storage reservoirs that feed both streams and groundwater.

Here I investigate the distinctly different water years of 2011 to 2013 and illustrate how sensitive subalpine meadow systems are to extreme variation in weather. Focused on three main themes of resiliency, thresholds and change in high elevation meadow ecosystems, I found that a) nonlinear responses in ecosystem processes can cause ecosystems to shift from a sink to a source of carbon contributing to local (degradation of ecosystem), regional (loss of ecosystem services) and global effects (feedback to atmosphere), b) winter drought and spring frost events can significantly damage vegetation, reducing the productivity of vegetation, which leads to a decrease in carbon sequestration, c) thresholds in moisture availability can influence the magnitude of carbon loss via soil respiration, d) Longer growing season with little summer precipitation contributed to severe desiccation of meadow soils which caused an irreversible change in the structure of the soil and subsequent loss of porosity and permeability. A new method was also developed to detect a signature of historic dryness in soils.

Findings of this study highlight the need to consider the timing of seasonality in mountain regions on the coupled hydro-biogeochemical dynamics when trying to assess resilience of meadow systems under a changing climate.

CHAPTER 1. Introduction

1.1 Background

High elevation wetland meadows in the Sierra Nevada of California have long been valued for not only their aesthetic importance on a landscape scale but also as hotspot of biodiversity in the mountains (Viers 2013). Recently though, their role in mountain hydrology has been taken center stage with the inclusion of the potential for restored and pristine meadows to contribute to water storage and water reliability for California in the California Water Plan Updates 2009 and 2013 (http://www.waterplan.water.ca.gov). While there has been a recent increase in our understanding of meadow hydrology in the Sierra Nevada (Hammersmark et al. 2008, Loheide II S.P. and Gorelick 2007, Loheide II S.P., and Lundquist, J.D. 2009, Lowry et al. 2010, McClymont et al. 2010), there has been only a limited amount of studies that have investigated biogeochemical cycling in theses systems (Saleska et al. 1999, Shaw and Harte 2001, Svejcar and Riegel 1998, Torn and Harte 1996), and almost no studies that focus on the physical characteristics of meadow soils and their relationship to past and present climates (Blank et al. 1995). If we are to attempt to quantify the value ecosystem services that meadows provide for California and include them in water planning, there is a great need for better understanding of how coupled hydrological and biogeochemical processes in meadows will respond to current climate threats.

High elevation meadows of the Sierra Nevada persist primarily due to the presence of a sustained high water table throughout much of the year that slows the rate of decomposition and allows for the accumulation of soil organic matter. During the past 10,000 years after the last Glacial Maximum, soils in these systems began to form, but it wasn't until around 4500 years before present that a shift to a moist hydroclimate allowed for peat soils to accumulate (Anderson and Smith 1994). Future changes in climate and or management practices that can increase the minimum depth to which the water table receeds in the growing season will presumably impact a range of ecosystem processes including carbon accumulation and loss. The magnitude of the impacts of present and future desiccation on meadow hydrological and biogeochemical resilience, thresholds and change is a critical missing component of meadow research.

1.2 Scope of the Dissertation

The overall theme of this dissertation is to enhance knowledge of the ability of high elevation meadows in the Sierra Nevada, California to respond to a changing climate through an examination of coupled hydrological and biogeochemical processes. The motivation for this study was the sequence of extreme years from 2011 to 2013 in the Central Sierra Nevada of California that caused a dramatic shift in the length of growing season and desiccation of meadow soil. This extreme seasonality is expected to have immediate impacts on meadow resiliency, through exceedance of hydrological thresholds resulting in changes in the rates of organic matter decomposition and soil structure.

1.3 Objectives of the Study

The aim of this dissertation was to increase understanding of coupled hydrological and biogeochemical processes (specifically carbon cycling) in high elevation meadows in response to extreme change in seasonality. The specific objectives addressed in this study are as follows:

1) To determine if extreme change in seasonality can cause an increase and/or decrease in ecosystem processes such as aboveground primary productivity and soil respiration to the extent that the ecosystem switches from a sink to a source of carbon.
2) To investigate the mechanisms that control organic matter decomposition in meadow soils subjected to a decrease in moisture availability.
3) To develop a method to induce soil consolidation through capillary suction which will help illuminate the soil response to desiccation in these systems.
4) To determine the extent to which desiccation of high elevation peat soils results in irreversible changes of soil structure and increased rates of organic matter decomposition with a focus on the impacts to the ability of these systems to provide ecosystem services.

1.4 Organization of the Dissertation

This dissertation focuses on the impact of a changing climate on high elevation meadow hydrology and carbon cycling. It is organized around the three main themes of resilience, thresholds and change. The remaining chapters are summarized below:

Chapter 2: Investigation on the impact of early spring and severe frost on the carbon cycling of high elevation meadows. Aboveground net primary productivity and surface carbon efflux were monitored over a three year period from two separate meadows at the crest of the Sierra Nevada Mountain Range to investigate how extreme changes in seasonality influence ecosystem processes in the meadow. We found that an early spring increases the likelihood of severe frost events that influence the overall productivity in the meadow. The date of snowmelt in 2012 and 2013 years occurred over 57 days earlier than in 2011 and resulted in multiple severe frost events that contributed to an average 39% reduction in aboveground biomass. In addition, dry conditions contributed to an average 100% increase in the cumulative soil respiration in the meadows. These results show how the depth and duration of snowpack can have profound impacts on meadow resilience and question how resilient the meadow systems are to prolonged change in seasonality.

Chapter 3: This chapter is a controlled laboratory investigation to better understand the response of high elevation meadow carbon mineralization to increased drying due to drought conditions. A long-term soil incubation was conducted utilizing soils from the Harvey Monroe Hall Research Natural Area. Soils from across the hydrologic gradient in the meadow and from three different depths were incubated for 392 days at five different water potentials. Gas samples taken at intervals over the incubation duration were analyzed for carbon dioxide in order to determine the effects of drying on carbon mineralization. We found that contrary to our initial hypothesis, carbon mineralization peaked at the wettest and the driest water potentials, indicating the presence of two separate pools of labile carbon that can be accessed only after a threshold of drying is reached in the soil. I propose a conceptual model of threshold-induced decomposition to explain the results.

Chapter 4: This chapter introduces a new method for measuring soil consolidation that occurs via desiccation induced capillary suction. We propose a modified triaxial system that combines the principles behind a triaxial system to measure volume change, a permeameter to measure hydraulic conductivity and a pressure plate apparatus to apply induce drainage of the sample. This system will simultaneously measure water drained from the sample and subsequent volume change as a consequence of desaturation. We can also reverse flow and measure saturated hydraulic conductivity before and after any drainage test.

Chapter 5: This chapter synthesizes the knowledge gained from this dissertation to better understand how extreme drought can cause immediate and irreversible change to both the hydrology and carbon cycling of the high elevation meadows. With the critical keystone role that meadows play in health of mountainous watersheds, in addition to their role in storing, filtering and releasing of water, we find that the exceedance of the historic limit of dryness in the meadow is a key element to determining their resilience to a changing climate. We find that minor exceedance of the historic limit of dryness can cause immediate and irreversible changes to soil structure, which will influence how and when water can move through the meadows. In addition, the initial drying and exceedance of historic limit can cause a large loss of carbon in the meadow, but further drying can be expected to reduce carbon mineralization. It is not until severe desiccation occurs that we can expect a secondary large pulse of carbon to be mineralized in the meadows.

Chapter 6: This chapter summarizes the knowledge gained from the previous chapters and highlights the need for continued research to better understand how a changing climate can modify resilience in high elevation meadows.

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Viers J., SE Purdy, RA Peek, A Fryjoff-Hung, NR Santos, JVE Katz, JD Emmons, DV Dolan, and SM Yarnell. 2013. Montane Meadows in the Sierra Nevada: Changing Hydroclimatic Conditions and Concepts for Vulnerability Assessment. University of California, Davis. Report no. CWS-2013-01.

CHAPTER 2. Early spring severe frost events and drought induce rapid carbon loss in high elevation meadows

Abstract

By the end of the 20th century, the onset of spring in the Sierra Nevada mountain range of California has occurred on average three weeks earlier than historic records. Superimposed on this trend is an increase in the presence of highly anomalous "extreme" years, where spring arrives either significantly late or early. Depending upon the year, nonlinear responses in ecosystem processes can cause these meadow systems to shift from a sink to a source of carbon contributing to local (meadow degradation), regional (loss of ecosystem services) and global effects (feedback to atmosphere). In this study, we assess the impact of extreme winter precipitation variation on aboveground net primary productivity and soil respiration over three years (2011 to 2013). We found that the duration of snow cover and presence of early spring frost events contributed to an average 100% increase in soil respiration in 2012 and 2103 and an average 39% decline in aboveground net primary productivity (ANPP) observed over the same years. The overall growing season length increased by 57 days in 2012 and 61 days in 2013. With the increase in growing season it was expected that ANPP would increase, but instead we found a dramatic drop in productivity that is attributed to early spring frost events and drying of meadow soils in both winter and summer. Our findings demonstrate that even small changes in climate can alter resiliency of these keystone ecosystems that depend on a stable climate.

2.1 Introduction

Magnitude and timing of extreme weather events have recently gained attention for their potential to alter ecosystem processes (Holmgren et al. 2006, Jentsch and Beierkuhnlein 2008, Jentsch et al. 2011, Walther et al. 2002) and concerns over the ability of natural ecosystems to respond to rapid changes (Craine et al. 2012). Extreme interannual change in weather (for example, from a very “wet” to a very “dry” year and vice versa) may trigger rapid carbon loss from an ecosystem (Galvagno et al. 2013, Reichstein et al. 2013). For high elevation mountain ecosystems in particular, the seasonal timing of the accumulation and melting of the snowpack is crucial for supplying abundant water to low-lying communities and high-elevation forests (Trujillo et al. 2012), and for promoting meadow productivity (Baptist et al. 2010) and soil carbon storage (Aurela et al. 2004, Galvagno et al. 2013). It is expected that earlier snowmelt will result in drying of meadow soils over the course of the growing season. This drying may lead to increased carbon storage through an increase in the net primary productivity of the system, but it can also lead to a loss of carbon through increased rates of decomposition. Whether the ecosystem remains a sink of carbon or shifts to a source of carbon will have large implications on the ability of the meadow to filter, store, and release water to the river systems. Prolonged conditions that result in a significant loss of carbon can eventually trigger a tipping point to a regime shift in the meadow.

The coupled hydrological and biogeochemical cycles in high elevation meadows are influenced by the depth and duration of the annual winter snowpack that acts as an insulating blanket during the winter (Inouye David W 2008). Not only does the winter snowpack protect the meadow soils from large temperature fluctuations and winter desiccation, it also functions to recharge the meadow soils during the spring snowmelt (Loheide II 2009). There is a dynamic two-way relationship between hydrology and soil organic matter (SOM) processes in such high elevation systems. Hydrology exerts a strong control on storage, stability, and composition of SOM (Heimann and Reichstein 2008) in the meadow soils while SOM dynamics controls the ability of the meadows to provide ecosystem services such as filtering, storing and releasing water to the river systems. The essential nature of those ecosystem services warrants a “keystone” status of mountain meadows in terms of hydrology. Watersheds that have lost meadow functioning due to degradation have limited water storage capacity and ability to attenuate floods (Brown 2013). This results in a flashy system, where the watershed responds rapidly to precipitation events. Furthermore, high-elevation meadows, which are hotspots of biodiversity (Myers et al. 2000) and function as breeding grounds for many organisms, are likely to be a key indicator of the overall health of central Sierra Nevada watersheds.

While interannual variations in snowpack depth and duration are normal in the Sierra Nevada (Kapnick and Hall 2009), consecutive years with extreme water conditions can significantly increase or decrease the overall length of the summer growing season and duration of snow free days, which will directly affect soil carbon storage. Previously, hydrologic modeling research in the Sierra Nevada has highlighted the sensitivity of that region's watersheds to earlier onset of spring and increased duration of low flows (Null et al. 2010). They have shown that some watersheds that are highly vulnerable to an increase in duration of low flows with climate warming, also contain the largest mountain meadow area. An increase in the duration of low flows can cause meadows systems to dry down significantly causing feedbacks to ecosystem processes. If the trend in the onset of spring (Cayan et al. 2001) continues, and meadows dry down earlier in the growing season, we can expect an increase in the decomposition of soil organic matter as the normally saturated soils become aerobic. This could also potentially impact river systems through a reduction in the ability of meadows to contribute to baseflow as they dry down. In addition, the timing of the onset of snowcover in the early winter can impact meadow soils and biota due to the widely fluctuating soil and air temperatures. The meadow soils in theses systems remain at 0°C as soon as snow accumulates in a continuous snowpack. This insulating layer protects overwintering biota and prevents drying of the meadow soil. Colder winter temperatures coupled to lack of continuous snowpack renders meadow soils and biota susceptible to severe desiccation which will impact ecosystem processes such as soil respiration and productivity in the following summer growing season.

An increase in interannual variation in the onset of spring has the potential to dramatically affect the balance between carbon storage (mainly through input of carbon from above and belowground biomass) (Chivers et al. 2009, Galvagno et al. 2013) and loss (through soil respiration and leaching) (Alm et al. 1999, Xiang et al. 2008) in meadow soils. A period of rapid carbon loss for the organic-rich high-elevation meadow soils can trigger a positive feedback loop that contributes to declining soil moisture (Orchard and Cook 1983), further organic matter decomposition and reduced plant productivity through changes in soil structure (Stephens et al. 1984) .

In mountain meadow ecosystems, changes in the timing of spring snowmelt have already been shown to influence plant phenology (Price and Waser 1998, Wipf et al. 2009), interactions between plants and pollinators (Thomson 2010), and longer term changes in meadow vegetation community structure (Forrest et al. 2010). In order to examine how the timing and duration of snow cover and presence of early season frost events and drought can influence ecosystem processes we monitored changes in surface carbon dioxide flux and aboveground net primary productivity over three consecutive summers (2011 to 2013) in two high elevation meadows in the Central Sierra Nevada mountain range of California.

2.2 Methods

2.2.1. Site description

Our study was conducted in two subalpine meadows with varying hydrologic regimes located at the crest of the Sierra Nevada mountain range along the boundary of Yosemite National Park (YNP) (Figure 2-1). Both meadows were formed as a direct result of past glaciation. Their resulting geomorphic position in the landscape remains conducive to high water tables throughout much of the growing season. One meadow is located in the Harvey Monroe Hall Research Natural Area (Hall RNA) at a 3200-m elevation on a large medial moraine on the eastern side of the central Sierra Nevada. The mean daily temperatures range from -4.9° C to 12.9° C (Taylor 1984). The soils in the Hall RNA are characterized as Inceptisols with the suborders Andic Cryumbrepts and Lithic Cryumbrepts (Staff).

For a contrasting type of meadow, we chose Dana Meadows, which is located at a 3000-m elevation along YNP's Tioga Pass Road in a U-shaped glacial valley with hummocky ablation till. Dana Meadows exhibits mean temperatures similar to those of the Hall RNA and an average precipitation of 1000 mm/year. The soils in Dana Meadows are classified as Inceptisols with the suborders Xeric Distrocryepts and Vitrandic Eutrocryepts (Staff).

2.2.2. Field methods

In July 2010, transects were established along a hydrologic gradient at two locations in the Hall RNA and two locations in Dana Meadows. The hydrologic gradient in the meadow was established using vegetation associations as a proxy for water table depth (Allen-Diaz 1991). In all four transects, the same vegetation type was utilized to identify each meadow region: *Carex filifolia* in the dry (xeric) sites, *Ptilagrostis kingii* in the intermediate (mesic) sites and *Carex sp.* in the wet (hydric) sites. Three replicate soil collars were inserted in the soil at depths of approximately 3-5 cm (the variation was due to differences in soil characteristics) at three hydrologically distinct regions of the transect (designated as: dry, intermediate, and wet). The collars are located approximately 2 meters apart. The soil carbon efflux was measured using a LI-COR 8100A portable infrared gas analyzer (LI-COR Biosciences, Lincoln, Nebraska USA), fitted with a portable 10-cm soil respiration chamber. After a 45-second pre-purge, one-minute measurements were recorded and were followed by a 30-second post-purge. Weekly measurements were recorded during the first half of the growing season, followed by biweekly measurements through September. All measurements were taken at mid-day from collars with vegetation left intact. In each of the 4 transects, there were 6 collars in 2011, and 18 collars in 2012 and 2013. Aboveground productivity was estimated by harvesting the total biomass in six 20 cm square quadrats in each region (dry, intermediate and wet) of the transect at peak production each year, oven dried till constant mass was recorded at 50°C, and weighed. Historical and current meteorological data, including snow water equivalent data were obtained from the California Department of Water Resources station for Dana Meadows (ID: DAN).

2.2.3. Satellite-based remote sensing imagery

The Terra/MODIS surface reflectance (MOD09Q1.5) 8-day L3 global 250-m product was downloaded directly from the Land Processes Distributed Active Archive Center of the United States Geological Survey (https://lpdaac.usgs.gov). This level 3 surface reflectance product, which had been radiometrically corrected and georeferenced, provided a measure of the surface reflectance at the ground level in the absence of atmospheric scattering or absorption. The data were projected in a custom sinusoidal projection specific to the MODIS imagery. The eight-day composite images represented the maximum surface reflectance value for that time period and minimized the impacts of clouds and aerosols.

2.2.4. Processing MODIS imagery to NDVI

In the first stage of processing, the MODIS product MOD09Q1.5 was reprojected from a custom sinusoidal projection to the California Albers projection. The latter is a version of the Albers Equal Area projection optimized for statewide calculations. Bands 1 (620-670 nm) and 2 (841-876 nm) were utilized to calculate the NDVI over the entire MODIS image. The following equation was used: NDVI = (band 2 – band 1) / (band 2 + band 1). The resulting NDVI product was resampled down to 30 m and was used to produce an average NDVI for the entire meadow polygon region. The meadow polygons, resulting derived data layer and the associated metadata are currently being prepared as a spatial data product by the U.S. Geological Survey's Western Ecological Research Center at the Yosemite Field Station (Berlow et al. 2013).

2.3 Statistical Analysis

Soil respiration rates for each collar were integrated over time to determine the cumulative CO_2 efflux for the growing season. Missing data was filled in via linear interpolation between the prior sampling date and the next date sampled. Repeated measures analysis of variance (RM-ANOVA) was utilized to determine significant differences between the effects of moisture class across the three years for both ANPP and cumulative CO_2 flux data. The different sites were utilized as replicates. If the RM-ANOVA model was significant, a Tukey's post hoc test ($p< 0.05$) was utilized to assess differences between means. In addition to determine the effect of year within a moisture region of the meadow, a subset of data was created and a one way RM-ANOVA model was utilized to determine significance within moisture classes across years. If the model was significant, a Tukey's post hoc ($p<0.05$) was used to determine differences between means. Data was tested for normality prior to analysis using the Shapiro-Wilks test. All statistical analyses were conducted using R statistical software (r-project.org).

2.4 Results and Discussion

The last several years in California have been marked by extreme seasonal weather on either ends of the spectrum. The 2011 water year (October 2010 through September 2011) was the seventh-wettest year on record (1929-2012) in YNP, with the April 1 snow water equivalent in Dana meadows 156% of the 50 year mean (1951-2000) (Table 2-1). The 2012 water year was the fifth-driest year on record with only 49% of the mean SWE and the 2013 water year ranked the driest year on record with 25% of the mean SWE (Table 2-1). Looking at the entire historic record of Dana Meadows SWE, there is an increase after 1969 in the number of years with SWE values greater or less than one standard deviation from the mean (Figure 2-2). This apparent increase in the SWE variability corresponds to trends found in increase in the variability of streamflow in Central California around the same time period (Pagano and Garen 2005). The current 2014 water year has continued the dry trend with little to no continuous snowpack in the Sierra Nevada at the end of January 2014.

In ecosystems dependent upon the enduring winter snowpack to insulate them from freezing events, the timing of the first day of continual snow cover for the winter can be critical to biological communities (Schimel et al. 2004). Likewise, the duration of that snow cover and timing of subsequent spring melt plays an essential role in microbial turnover (Lipson et al. 2002, Nemergut et al. 2005), plant phenology (Walker et al. 1995), and meadow hydrology (Bales et al. 2006, Jordan 1978). Not only was the depth of snowpack distinctly different in the three consecutive years, but also the duration of snow cover differed greatly in all three years (Figure 2-3). The water year 2012 was especially anomalous with no continual snow cover until mid January. A significant ice storm occurred over the bare soils on December 4, 2012 with widespread needle damage to conifers noted at the study site after snowmelt. With the exception of January, the average maximum air temperature in Dana Meadows was warmer in 2013 and 2012 through July of each year, as compared to 2011, with the spring (Mar-May) mean temperature increasing by 2-3 °C (Table 2-2). This early warming is a contributing factor to the onset of an early spring in those years. The mean monthly minimum temperatures show warmer spring and summer temperatures in 2012 and 2013 as compared to 2011 (Table 2-3).

2.4.1. Ecosystem Response

The extreme seasonal changes from 2011 and 2012/2013 caused dramatic shifts in the onset of spring and in the number of snow-free days in the meadows of YNP. In Dana Meadows, the first snow-free day in 2012 and 2013 occurred 57-61 days earlier than in 2011, and the growing season increased by 35-37%, from approximately 106 days in 2011 to 163/167 days in 2012/2013. The documented shift to an earlier onset of spring in the Sierra Nevada (Cayan et al. 2001) appears to have altered the response of the meadow ecosystems; rather than increasing their productivity, the earlier spring has rendered the meadows more sensitive to late winter/early spring frost events (Forrest et al. 2010, Inouye DW 2001, Inouye David W 2008, Inouye D.W. et al. 2002).

The maximum and minimum temperatures in the meadow show a clear seasonal trend, with very few frost events occurring within a normal growing season (Figure 2-4). The seasonal trends are similar between years, but when the growing season is initiated is critical for assessing potential frost impacts on newly sprouting vegetation. As the snow melts, it causes saturation of the meadow soil and plants respond rapidly to this moisture and available nutrients by sending up green shoots. This leaves them susceptible to freezing temperatures. If spring arrives earlier, as in 2012 and 2013, there is an increased likelihood of a severe frost event to damage newly sprouting vegetation. This pattern was evident during two frost events that occurred in 2012 after the snow had cleared from the meadow (Figure 2-4 and Figure 2-5). The first event occurred over a four-day period that peaked on May 27, when the temperature dropped to -10°C. The second event occurred over three days beginning on June 5 that included a low temperature of -9.4°C. Because meadows undergo a rapid greening within a week of snowmelt, frost can damage sensitive meadow species and reduce overall productivity (Inouye D.W. et al. 2002, Neuner and Hacker 2012). In 2013, there were several frost events that occurred on May 19 and 20 and on May 22 and 23 with a low at -5.5°C. Tranquillini (1964) has shown that high elevation plants are very frost resistant, however notes that plants dependent on an insulating snowpack are susceptible to frost damage even in minor frost events (Tranquillini 1964). Frost damage to vegetation was apparent on a larger scale using satellite-derived Normalized Difference Vegetation Index (NDVI) mapping. The NDVI time series for the three years is shown in Figure 2-3. A rapid greening was apparent at the beginning of the growing season for both years; however, instead of reaching a peak in 2012, the NDVI plateaued before the vegetation senesced in mid-summer, which indicates that the meadow vegetation was stressed and never reached maximum greenness in 2012. Normal meadow NDVI ranges from 0 during snowcover to a maximum of 0.45-0.65 at peak production and then falls to around 0.3 during senescence. In 2011, the peak NDVI occurred around 0.45, but in 2012 and 2013, the peak NDVI was just above the senescence value. Since the meadow soils begin the season saturated due to snowmelt and dry down over the growing season, it is unlikely that the plateau was caused by the meadows drying earlier in the growing season. If this were the case, we would expect to see a peak NDVI soon after the rapid increase at the beginning of the season. This is due to the fact that even in a dry year, the meadow will begin the growing season at saturation to near saturation. It may dry down earlier, but there would still be a maximum NDVI value from the growing vegetation. The pattern in 2013 is similar, though there is a small peak in early season NDVI, but the overall pattern is much lower than in 2011, indicating that although aboveground productivity did recover slightly from 2012, the meadow was still in a stressed state. There is an anomalous peak of NDVI that occurs between November 2011 and January 2012. This was the value of the senesced vegetation that lacked a continuous snowpack until mid January 2012.

The prolonged growing season in 2012 and 2013 should have lead to increased plant productivity if water and temperature were not limiting, but the productivity actually declined by an average of 39% in all the regions of the meadow (Figure 2-6). Averaged across all sites, the drop in productivity from 2011 to 2012 was the most significant of all three years ($p<0.000001$). The aboveground productivity in 2013 was still significantly lower than 2011 ($p<0.0001$) and didn't significantly increase from 2012, indicating that the system is still at a stressed state and not recovering rapidly. However, since belowground biomass was not quantified in this study, it is unknown if the meadow systems are adapting to the change in seasonality by putting more energy into belowground biomass. However, due the fact that meadows already allocate a greater proportion of their carbon inputs (60-80%) to roots, it is unlikely that a shift to more belowground production could offset the carbon losses via soil respiration in 2012 and 2013. There is a delicate balance between productivity and carbon loss via respiration and unless the plants drastically increased the belowground biomass in response to the early spring, the losses via soil respiration will override the carbon inputs from the additional belowground biomass.

Different hydrologic regions of the meadows (dry – wet) responded differently to the change in duration and amount of snow cover. There was no significant difference in the productivity of the dry regions of the meadow across the three years. However, both the intermediate and wet regions showed significant reduction in biomass from 2011 – 2012 (intermediate, $p<.0001$; wet, $p<0.0001$). Both regions of the meadow still had significantly lower productivity in 2013 than 2011 (intermediate, $p<0.0001$; wet, $p<0.0001$), but no significant change from 2012 to 2013.

The mean cumulative carbon flux shows an increasing trend over time (Figure 2-7). Averaged across all moisture regions of the meadows, the mean cumulative carbon flux from the meadows was significant with respect to year (RM ANOVA model; year, $p< 0.00001$, moisture, $p< 0.00001$, moisture:year, $p= 0.447$). However, this was driven mainly by the significance between 2011-2012 ($p< 0.00001$) and 2011-2013 ($p< 0.00001$). While 2013 continued to have larger cumulative fluxes in all regions of the meadow, they were not significantly different than 2012 ($p=0.481$). There was no interaction between moisture region and year. Comparing the individual moisture regions of the meadow, there was significant change in the cumulative carbon flux of the dry, intermediate and wet regions of the meadow from 2011 to 2012 and 2011 to 2013, but no significant change from 2012 to 2013 (Table 2-4).

This result suggests that easily decomposable soil organic matter, previously protected by a high water table, rapidly decomposed with the shift in environmental conditions in 2012 and 2013 (Berhe et al. 2012, Davidson and Janssens 2006, Laiho 2006, Waldrop et al. 2010). The short growing season in 2011 and wet conditions throughout the growing season effectively reduced the overall soil carbon efflux in all regions of the meadow. Although the 2011 water year was extremely wet, and the meadows experienced little drying, this had no apparent effect on the available moisture in 2012. One explanation for this finding may be the dryness caused by a lack of snowpack in December 2011 through mid-January 2012; when the snow finally accumulated in January, the meadows were extremely dry underneath the snowpack. These dry soil conditions at the start of the 2012 growing season and below average snowpack coupled to above average surface temperatures throughout the 2012 growing season led to the extreme drying of the meadow soils and subsequently large carbon fluxes. The 2013 water year was the driest on record for Dana Meadows and there was a continuing trend of high mean cumulative carbon flux from the meadows. Although the cumulative flux was similar in both 2012 and 2013, there was a difference in the timing of peak soil carbon efflux. In 2013, the peak soil carbon efflux occurred early in the growing season and rapidly declined in all regions of the meadow, whereas in 2012, the peak occurred in the middle of the growing season (Figure 2-8). This shift indicates how responsive these ecosystems are to seasonal variation. The sustained magnitude of the cumulative carbon flux from the system over two summers has resulted in a loss of over 6% of the total carbon stock in the meadows we studied.

With climate extremes occurring at an increasing frequency around the world, our data demonstrate that sensitive ecosystems respond rapidly to the changes in seasonality and may reach a tipping point sooner rather than later. Multiple years of ecosystem stresses such as frost or drought can potentially cause a regime shift in vegetation with ramifications to the cycling of carbon in these systems. The magnitude of loss was significant given the small areal extent of these meadows, which is not proportional to their importance to overall ecosystem functioning and keystone position on the landscape. If the frequency of extreme events continues in this region, coupled to a decline in meadow aboveground productivity, we can expect carbon stocks in the meadows to rapidly decline, leading to meadow degradation and a reduction in ecosystem services in these watersheds.

2.5 Acknowledgements

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2.6 Figures

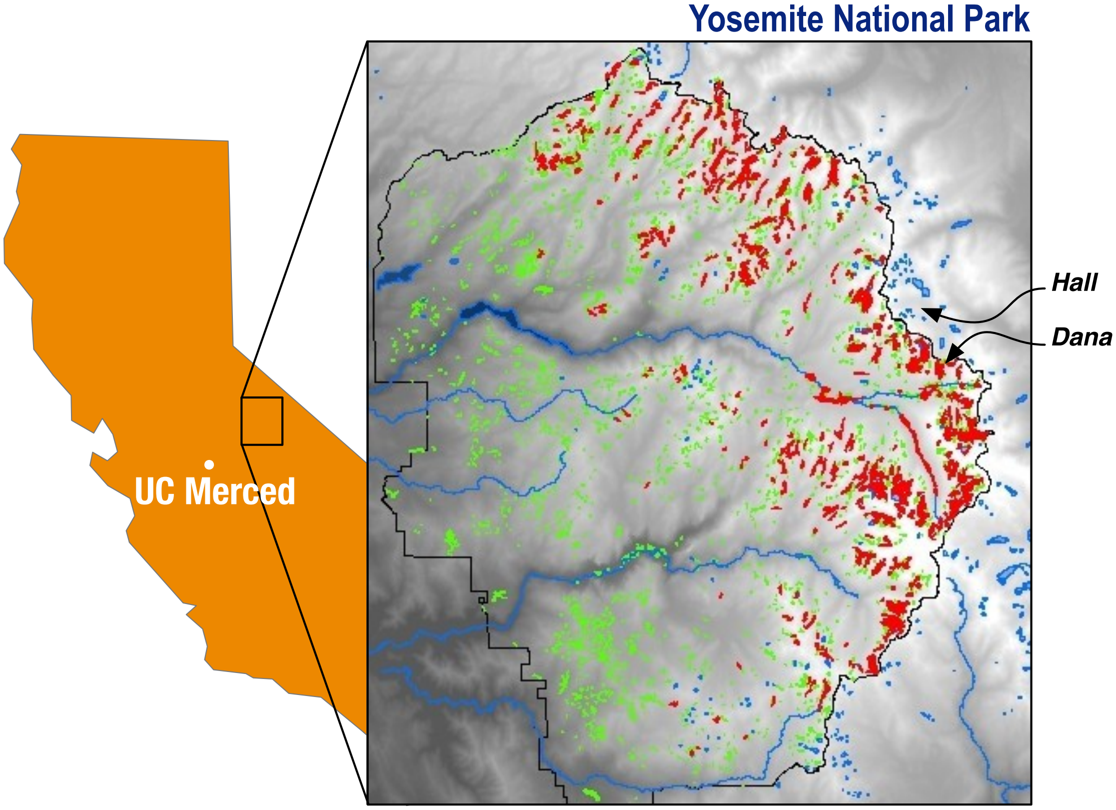

Figure 2-1 Map of study sites along the boundary of Yosemite National Park, California. Polygons represent the extent of meadow area in Yosemite National Park, with subalpine meadows (> 3 hectares and between 2600-3200 m) highlighted in red.

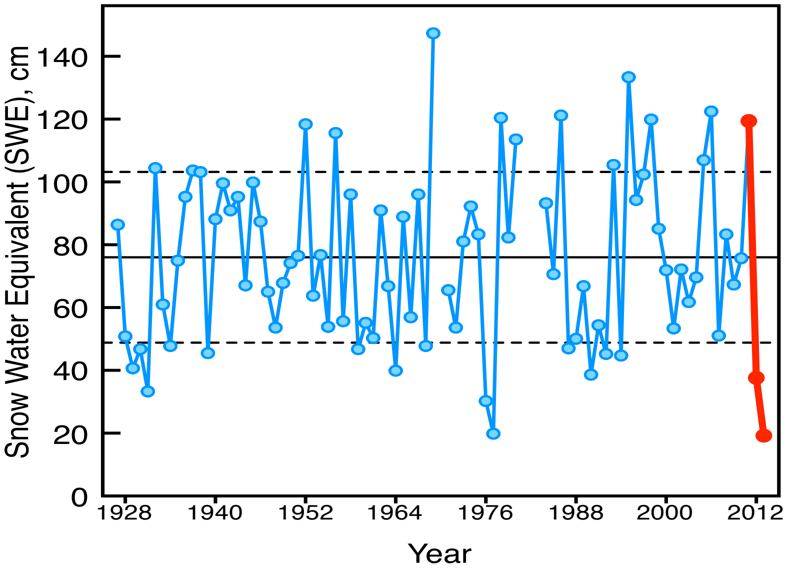

Figure 2-2 Historic SWE record for Dana meadows (Yosemite National Park) with 2011-2013 highlighted in red .

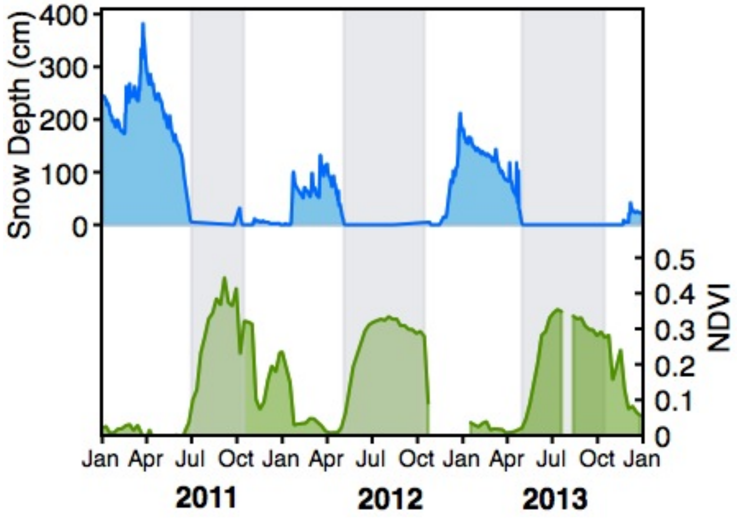

Figure 2-3 Snow depth (top panel) from Dana Meadows and NDVI (weighted mean average of all subalpine meadows in Yosemite National Park). Gray panels denote the growing season in the meadows as defined by the first day the meadow is snow free and the date where the NDVI crosses a threshold of 0.3.

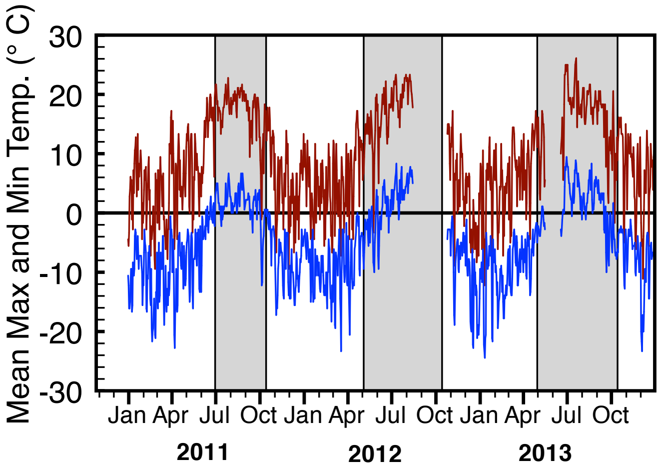

Figure 2-4 Maximum (red) and minimum (blue) daily air temperatures for Dana Meadow in 2011, 2012 and 2013.). Gray panels denote the growing season in the meadow as defined by the first day the meadow is snow free and the date where the NDVI crosses a threshold of 0.3.

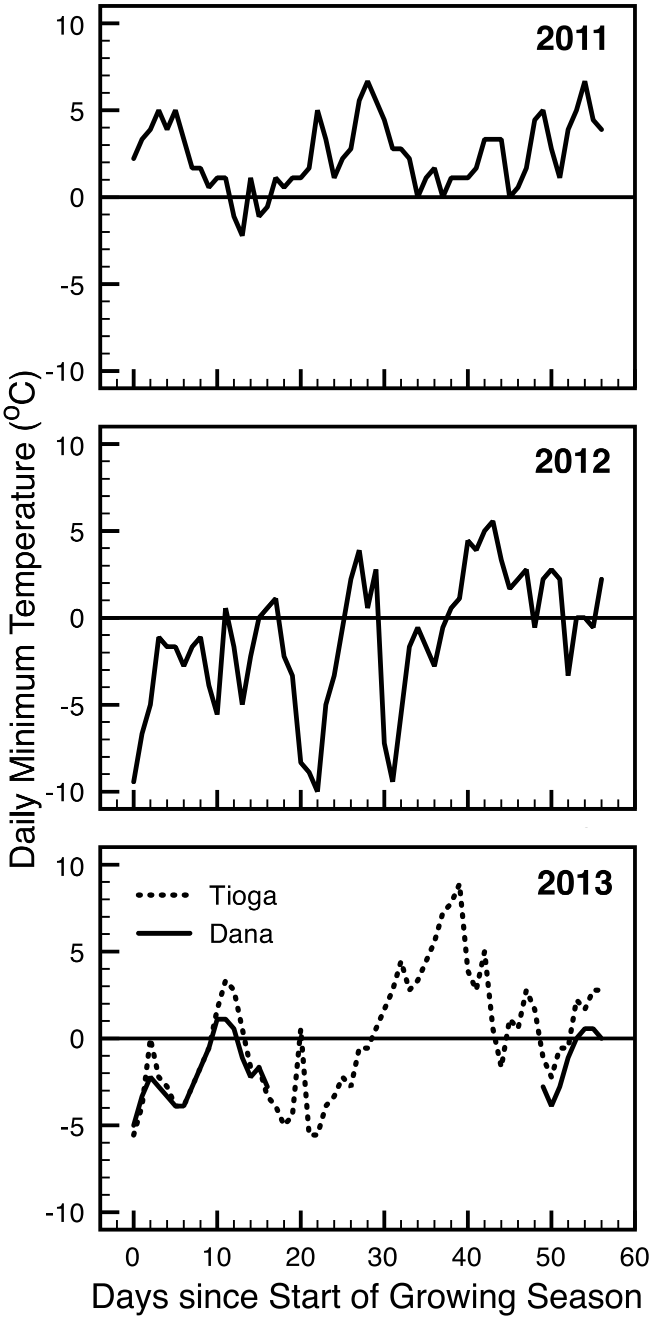

Figure 2-5 Daily minimum temperatures were used to compare frost events in the first sixty days since the start of the growing season (2011-2013) in Dana Meadow. Growing season was determined by the first day that the meadow was snow free each year. Dotted line represents data from a nearby meteorological station (Station id:TES) at north end of meadow was used for missing data in 2013.

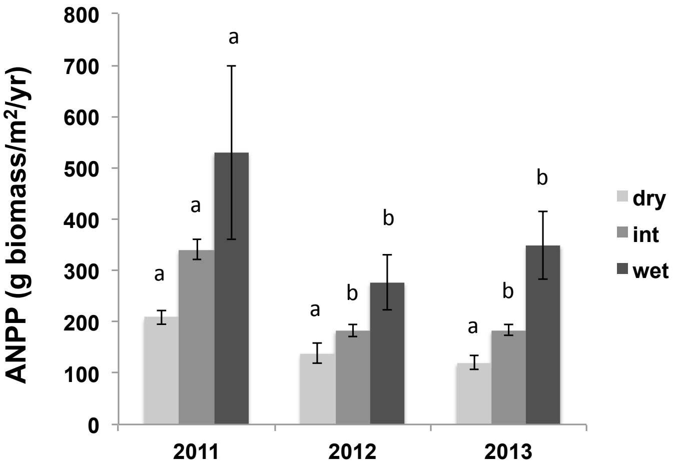

Figure 2-6 Mean annual aboveground net primary productivity averaged across all four sites for each moisture region in 2011, 2012, and 2013. Error bars represent standard error among sites. Letters denote significant differences in homogenous groups across years as determined by a Tukey post hoc test ($p<0.05$).

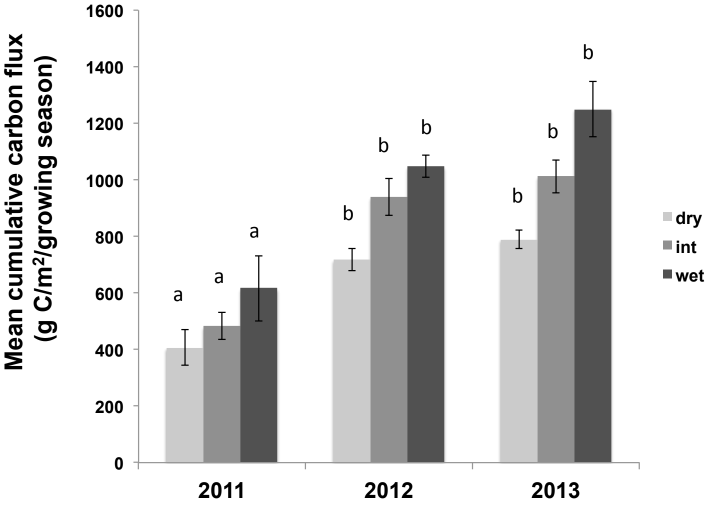

Figure 2-7 Mean cumulative carbon flux averaged across all four sites for each moisture region in 2011, 2012, and 2013. Error bars represent standard error among sites. Letters denote significant differences in homogenous groups across years as determined by a Tukey post hoc test ($p<0.05$).

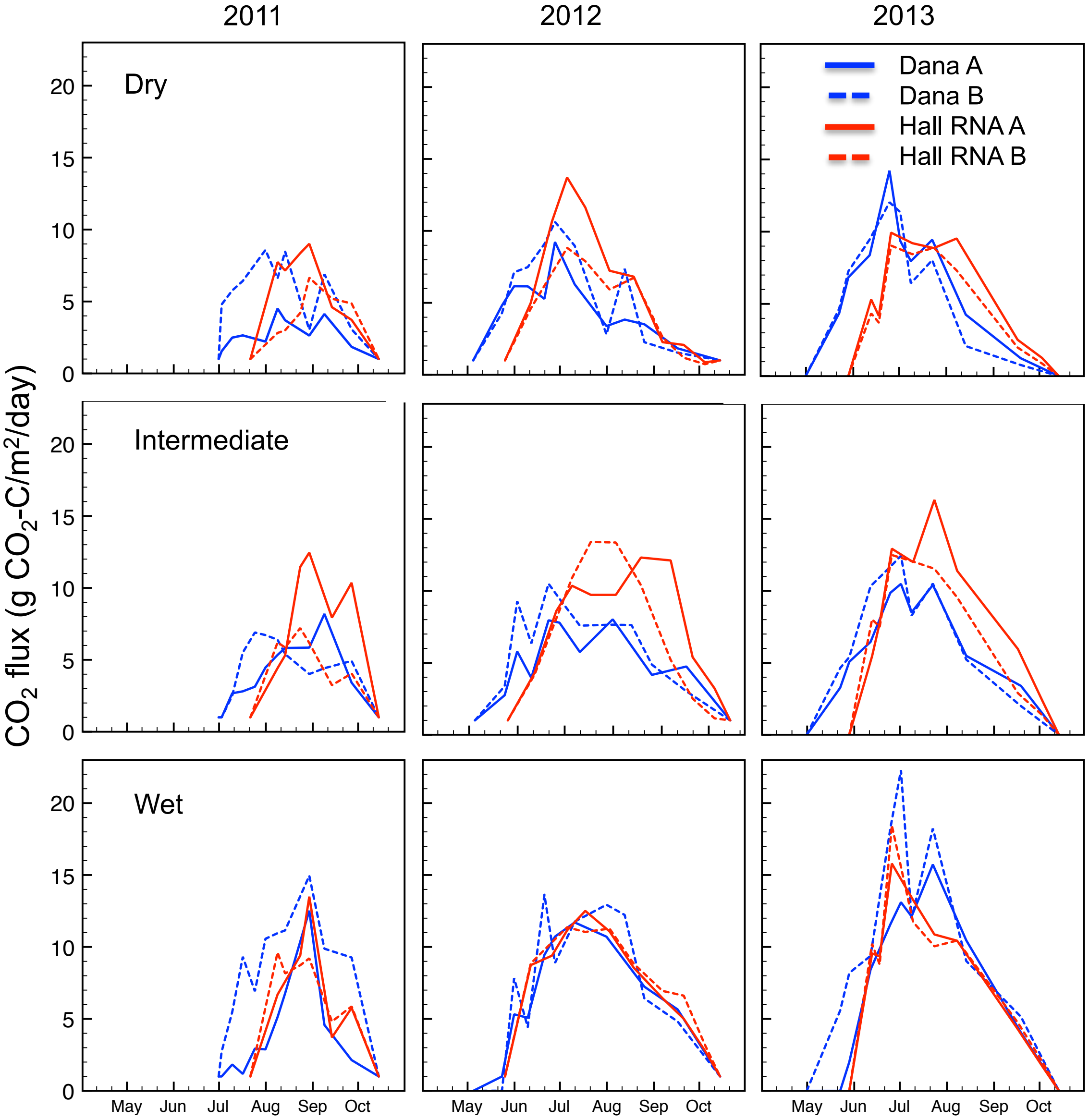

Figure 2-8 LI-COR surface CO2 flux data for 2011, 2012, and 2013 in Dana Meadows and the Hall RNA.

Table 2-1. Historic ranking of SWE in Dana Meadow

The top 12 driest years on record (1927-2012)				**The top 12 wettest years on record (1927-2012)**			
Rank	**Year**	**SWE (cm)**	**% of average**	**Rank**	**Year**	**SWE (cm)**	**% of average**
1	**2013**	**19.2**	**25**	1	1969	147.32	192.04
2	1977	19.81	25.83	2	1995	133.35	173.83
3	1976	30.23	39.40	3	2006	122.43	159.59
4	1931	33.27	43.37	4	1986	121.16	157.94
5	**2012**	**37.59**	**49.00**	5	1978	120.40	156.94
6	1990	38.61	50.33	6	1998	119.89	156.28
7	1964	39.88	51.98	**7**	**2011**	**119.38**	**155.62**
8	1929	40.64	52.98	8	1952	118.36	154.29
9	1994	44.70	58.27	9	1956	115.57	150.65
10	1992	45.21	58.94	10	1980	113.54	148.00
11	1939	45.47	59.28	11	2005	106.93	139.39
12	1930	46.74	60.93	12	1993	105.41	137.41

Table 2-2 Mean maximum air temperature (°C) in Dana Meadow. Mean seasonal temperatures (DJF, MAM, JJA, SON) are highlighted in gray. Historic mean maximum air temperatures from 2000-2010 are referenced in far right column.

	2013		**2012**		**2011**		**2010**	**Historic Mean (2000-2010)**
JAN	3.12		4.78		4.93			2.21
FEB	3.21	1.82	2.82	3.38	0.89	2.53		2.44
MAR	6.81		3.64		2.47			5.35
APR	9.11		8.06		5.78			6.62
MAY	10.75	8.89	11.86	7.85	7.67	5.31		11.24
JUN	15.97		16.28		13.48			15.44
JUL	20.05		19.87		18.39			19.62
AUG	18.57	18.20	21.33	19.16	19.48	17.12		18.72
SEPT	15.11				17.17			14.59
OCT	10.43		13.24		11.94			7.06
NOV	7.00	10.85	6.17	9.70	3.68	10.93		5.04
DEC	4.68		-0.88		2.53		1.76	1.83

Table 2-3 Mean minimum air temperature (°C) in Dana Meadow. Mean seasonal temperatures ((DJF, MAM, JJA, SON) are highlighted in gray. Historic mean minimum air temperatures from 2000-2010 are referenced in far right column.

	2013		**2012**		**2011**		**2010**	**Historic Mean (2000-2010)**
JAN	-11.20		-7.85		-9.48			-10.92
FEB	-12.64	-12.12	-11.07	-9.69	-12.57	-10.45		-11.40
MAR	-8.01		-9.57		-11.00			-10.11
APR	-6.33		-7.18		-9.54			-8.56
MAY	-2.03	-5.46	-3.39	-6.71	-7.35	-9.30		-3.75
JUN	1.53		0.44		-2.20			0.34
JUL	4.87		3.46		2.29			4.25
AUG	2.85	3.08	5.89	3.26	2.76	0.95		2.84
SEPT	0.17				1.56			-0.93
OCT	-4.89		-3.06		-2.62			-4.59
NOV	-5.78	-3.50	-6.36	-4.71	-8.43	-3.16		-7.51
DEC	-10.09		-12.51		-10.16		-9.28	-10.37

Table 2-4 One-Way RM-ANOVA results from homogeneous moisture regions of the meadow.

	Dry	**Intermediate**	**Wet**
2011-2012	P< 0.003	P< 0.0007	P< 0.006
2011-2013	P< 0.001	P< 0.0004	P< 0.002
2012-2013	P= 0.798	P= 0.888	P= 0.667

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CHAPTER 3. Soil organic matter decomposition in high elevation wetlands is driven by hydrological thresholds

3.1 ABSTRACT

While peatlands are garnering much attention for their greenhouse gas feedback potential in a warming climate, the coupled biogeochemical and hydrological impact of structural and physical changes in these types of systems as a result of desiccation has not been studied in detail. The cyclic drawdown/recharge of the water table that exists in most peatland systems impose important controls on organic matter storage and decomposition as well as soil physical properties. The normally saturated soils function to reduce the aerobic decomposition and allow organic matter to accumulate over time, which increases the ability of these systems to retain water. However, during dry years, there is potential for peatland systems to dessicate, allowing decomposition to proceed at a faster rate than normal. In order to better understand how high elevation peatlands will respond to increasingly dry years, we incubated meadow soils collected along a hydrologic gradient at 5 different water potentials and measured the CO_2 flux at intervals for over one year to determine if desiccation of meadow soils influences cumulative carbon flux. Contrary to what was expected, we found that the cumulative carbon mineralization was greatest at the highest (0.1 bar) and lowest (4 bar) water potential, across all regions of the meadow. We propose a conceptual model that illustrates how structural changes in organic soils during drydown can explain our results.

3.2 Introduction

High elevation meadow ecosystems in snowmelt-dominated watersheds play a fundamental role in carbon sequestration as well as in filtration, storage and slow release of water to streams. The hydrology and biogeochemical cycling of essential elements in such ecosystems are inextricably linked. It is very well documented that the concentration of soil organic matter (SOM) has a strong influence on water holding capacity of soil (Boelter 1969, Hudson 1994, Rawls et al. 2003) and contributes to the essential ecosystem service of water storage in high elevation meadows (Hammersmark et al. 2008, Ratliff 1985). Similarly, there is a growing body of literature that discusses the influence of soil water content on rate of organic matter (OM) loss and storage in soil (Davidson et al. 2000). However, relatively less is currently known about how the response of OM decomposition in soils varies along a continuum of water potential, for soils that vary in availability, chemistry and distribution of OM in the soil matrix. We especially have poor understanding of water content and water retention thresholds that are critical in regulating rate of organic matter decomposition. Due to the significant influence of hydrology on maintenance of meadow ecosystem health (Hammersmark et al. 2008, Loheide II et al. 2009), such a missing link could prove critical for improving our understanding of mechanisms behind hydrologic resiliency of high elevation meadows and how changes in management practices or climate can impact that resiliency.

Early research on the dynamic and two-way relationship of SOM and water dynamics in soils showed conflicting results. Some early studies showed that SOM content had very little to no influence on water retention in soil (Feustel and Byers 1936), while others showed significant influence that was subsequently determined to apply only to coarse soils (Jamison and Kroth 1958). The influence of soil texture on the water holding capacity of soils was deemed more important than the quantity of SOM, due to the observed influence of SOM on coarse soils and not on fine textured soils (Bauer and Black 1992). This viewpoint has changed over time with research finding evidence that quantity of SOM influences the structure, and water holding capacity of a variety of soils (Rawls et al. 2003). Furthermore, although the recognition for impact of SOM on soil structure has been growing, there is still lack of sufficient studies that investigate the mechanisms by which SOM influences pore structure in the soil (Oades 1984), and the indirect effect hydrology affects organic matter (OM) dynamics by mediating soil structure (Kay et al. 1997). Systematic investigation of the relationship between soil water retention and SOM dynamics can help illuminate the mechanism behind the water holding capacity of high elevation meadows.

Better understanding of the relationship between OM biogeochemistry and hydrology in high altitude (or high latitude) systems that are typically characterized by a long wet and cold season is crucial for our understanding of how these systems are likely to respond to anticipated future climate change. In the Sierra Nevada Mountain range, high elevation wetland meadows tend to have high water table throughout the growing season (Ratliff 1985). In these ecosystems there is a natural drawdown of the water table during the summer months to within a meter of the surface. The accumulation of snowpack during the winter months and subsequent snowmelt functions to recharge the meadow soils, restoring the water table in the meadow (Wood 1975). The cyclic drawdown/recharge of the meadow water table has implications for regulating aerobic decomposition in the meadow (Chimner and Cooper 2003, Ise et al. 2008). The soils in these ecosystems are fairly young, having transitioned to the accumulation of peat around 4500 years before present (Anderson and Smith 1994).

Similar to boreal peatlands, the accumulation of carbon in soils of high elevation mountain ecosystems in the Sierra Nevada and other mountain ranges is primarily driven by low rates of decomposition as the soils stay near saturation for the bulk of the growing season (Clymo RS 1965, Malmer 1986), making these ecosystems a net sink of carbon dioxide from the atmosphere. While increasing temperatures as a function of a changing climate have implications for increasing decomposition of soil organic matter in many ecosystems, the perturbations on the hydrologic cycle can have a more dramatic impact on whether these systems are a sink or source of carbon in any given year (Funk et al. 1994, Moore and Knowles 1989).

Studies in the laboratory have shown that a lowering of the water table can stimulate carbon dioxide flux from peat microcosms (Blodau et al. 2004, Funk et al. 1994, Moore and Dalva 1993). However field studies have shown mixed results with some reporting no significant change in soil respiration (Freeman et al. 1996), and yet others reporting importance of microtopographical controls (Strack and Waddington 2007), and even significant carbon losses (Oechel et al. 1998) . In one study, field based measurements of ecosystem respiration showed little to no response to drying whereas laboratory incubated peat cores showed a decrease in respiration with drying at the surface (Lafleur et al. 2005). Chivers et al found that drying in an Alaskan fen induced plant stress which was more responsible for reducing the carbon sink in their ecosystem through a reduction of gross primary productivity rather than an increase in ecosystem respiration (Chivers et al. 2009). This disagreement warrants an explanation and a systematic study to evaluate how moisture regulates decomposition in such systems.

So far, an overwhelming majority of the studies that have investigated the effect of drying on decomposition of organic matter have focused on the biogeochemical or microbial response to one or several drying events. However, in highly organic soils, changes in the physical structure of the soil as drying proceeds may provide some insight on whether the system loses a lot of C through gaseous as well as dissolved fluxes of carbon. Peat soils have a high proportion of macropores, and a high capacity to retain water even through a dry period (Beven and Germann 1982). However, soil consolidation and subsequent shrinkage as a result of drying can reduce the proportion of large pores and create a wetting inhibition (Schwarzel et al. 2002). This in turn has implications for carbon dioxide dynamics in peat soils with changes in soil structure influencing the redistribution of soil moisture via capillary rise (Wessolek et al. 2002). Wessolek et al. (2002) found that carbon dioxide losses in a peat soil could be linked to the ability of a soil to move water via capillarity. This may help explain previous disparate results in the field and laboratory that didn't take into account the linkage between pore structure and water dynamics in relation to carbon dioxide production. This can be due to the fact that the shrinkage of pores in soils is dependent upon the maximum intensity of previous drying cycles (Peng et al. 2007). It is plausible that future drying can potentially alter carbon cycling in hydrologically sensitive ecosystems.

In order to examine how drying events will influence mineralization of carbon in high elevation wetlands, we conducted a long-term soil incubation with samples drained to five different water potentials. We hypothesized that aerobic decomposition of soil organic matter would increase with decreasing (drier) water potentials until a critical water potential was reached, upon which a significant decline in aerobic decomposition was expected.

3.3 METHODS

3.3.1. Site description

Soils for the incubation were collected from a meadow at 3200 m elevation in the Harvey Monroe Hall Research Natural Area (Hall RNA). The Hall RNA is located at the crest of the Sierra Nevada Mountains, in the Inyo National Forest, on the eastern side of the Sierras adjacent to Yosemite National Park. It is characterized by a cool climate with cold wet winters and cool dry summers. Mean daily temperatures range from -4.9°C to 12.9°C and average precipitation is 642 mm/year (Taylor 1984). The soils in the Hall RNA are characterized as Inceptisols with suborders Andic Cryumbrepts and Lithic Cryumbrepts.

Meadows in the high elevation ecosystems in the Sierras are characterized by hydrologic gradients that are defined by the geomorphology of the landscape. In the fall of 2011, bulk meadow soils were collected from across the hydrologic gradient in the study meadow. The hydrologic gradient in the meadows was delineated by vegetation (Allen-Diaz 1991), with dry meadows associated with Carex *filifolia*, intermediate meadows with Ptilagrostis *kingii* and wet meadows associated with Carex *vesicaria/utriculata* and Carex *scopulorum*. At ten random locations in each of the wet, intermediate and dry regions of the meadow, bulk soil samples were collected from the following three depths: 0-10 cm, 10-20 cm, and 20-40 cm (Figure 3-1). At the dry meadow locations, the depth of the lower soil sample only extended to 30 cm. These soils were packed on ice and transported to the laboratory at the University of California, Merced, where they were composited by location (dry, intermediate, wet) and depth, sieved to 8 mm to remove large roots and gravels and repacked to native bulk density in 6.35 cm tall x 5.08 cm wide plastic tubes. The repacked soil cores were then allowed to saturate from below overnight and weighed.

3.3.2. CO_2 measurements

Three replicate cores of each layer (n=135) were drained respectively to one of the following water potentials (ψ = 0.1, 0.3, 1.0, 2.5 or 4 bars) on a pressure plate apparatus following methods of Klute (2003). These water potentials were chosen to represent a range of potential dryness levels that could be expected in these soils. The dry meadow regions may experience dryness of 4 bar during a really dry summer, but this level would not be experienced in the intermediate and wet meadow regions in a normal year. It would take severe desiccation for 4 bar to be reached in those regions of the meadow. These repacked cores were then placed in one-liter mason jars, covered with plastic wrap and incubated at 20°C in the dark for over one year. The jars were weighed weekly and water was added to maintain the initial drained weight. Gas samples were pulled at the following intervals (day 1, 3, 7, 14, 28, 42, 63, 99, 140, 248, 392) by capping the mason jars for 24 hours and extracting 15 ml of gas through a septa inserted in the jar lid. These samples were then analyzed on a gas chromatograph, (Shimadzu GC2014) fitted with a thermal conductivity detector for determination of carbon dioxide concentrations.

3.3.3. Soil properties

Soil texture was analyzed at the University of California Analytical Laboratory by the hydrometer method (Sheldrick and Wang 1993). Soil pH was analyzed on three separate replicates in a 1:1 water dilution (Thomas et al. 1996). Percent carbon and nitrogen were analyzed in duplicate on ground bulk soil samples from each location and depth in the meadow on a Costech Elemental Analyzer.

At the same time that the bulk soils were collected, we extracted two intact and paired soil cores from the same three depths in the locations directly adjacent to where the bulk samples were collected. One sample core of each pair was saturated overnight and drained to 0.3 bar on the pressure plate apparatus. The final drained weight was measured and the cores were incubated in one-liter mason jars for 392 days with the rest of the composited samples. Water was added each week to maintain the constant drained weight. Gas samples were drawn at intervals for over one year at the same time as the composited cores. The second set of cores were saturated overnight in the laboratory and drained on a pressure plate apparatus over a month long period to determine the water retention curve for the samples. The resulting water retention data was fitted with Durner-van Genuchten water retention model (Durner 1994) van Genuchten (1980), which is appropriate for structured soils that exhibit bi-modal pore size distribution,.

$$\Theta = \sum_{i=1}^{2}(1 + (\alpha_i \psi)^{n_i})^{-m_i} \tag{1}$$

where $i = 1$ and $i = 2$ represent the macro- and micro-pores, respectively; and α , n and $m = 1 - 1/n$ are empirical shape parameters. The effective saturation $\Theta = (\theta - \theta_r)/(\theta_s - \theta_r)$ depends on the saturated water content (θ_s) and residual water content (θ_r). Bulk density values were estimated from the water retention curve data.

3.3.4. Data Analysis

In order to determine the initial pool of labile carbon available for decomposition, the cumulative carbon mineralization data was fitted with an exponential rise to max curve (Equation 2). This one pool model carries the assumption that all of the carbon in the sample decays at the same rate. A two-pool model was tested on the data, but since the majority of data could be fitted with the one pool model, we chose to utilize the one pool model for all the samples.

$$C_t = C_o(1 - e^{-kt}) \tag{2}$$

where, C_t is the cumulative mineralization of carbon after time (t); C_o is the pool of potentially mineralizable carbon in the sample; k is the rate of carbon mineralization; and t is the total time of the incubation

A one-way ANOVA, was performed to compare the difference between means rate of carbon dioxide mineralization across water potentials. If the model was significant ($p< 0.1$), a Tukey HSD post hoc test was performed in order to determine which treatments were significant. All statistical analysis was done on the R statistical software (rproject.org).

3.4 Results

3.4.1. Soil Properties

The soils in this region were classified as a sandy loam, with high percentage of organic matter. The percent carbon in top 0-10 cm depth ranged from approximately 6% to 34% across the hydrologic gradient (dry to wet) in the meadow. While the percent carbon below 10 cm rapidly dropped to 2-3% carbon in the dry and intermediate meadow regions, the wet meadow region maintained high (10-12%) carbon values at similar depths (Table 3-1). Bulk density ranged from 0.89 to 0.17 g/cm^3 moving from dry to wet across the meadow in the top 10 cm of the soil. The bulk density of the wet meadow soil increased to a maximum of 0.54 g/cm^3 at a depth of 40 cm, compared to approximately 1.56 g/cm^3 and 1.25 g/cm^3 in the dry and intermediate regions respectively. The soils are highly to moderately acidic with pH values ranging from 3.8 to 5.7.

An examination of the water retention curves from the series of intact cores showed that there is an initial large drop in water content as the soils progress from saturation to 0.05 bar (Figure 3-2). After the large pores are rapidly drained by gravity, the subsequent release of water by these sandy organic soils became gradual as the drying progressed. There was a secondary pulse of water released as the soils experience further drying at 4.0 bar.

3.4.2. Carbon mineralization

Overall, we found the largest carbon loss in the top 0-10 cm of the wet region of the meadow (between 30-40 g CO_2-C/g soil) at all water potentials (Figure 3-3). The intermediate and dry meadow regions also showed the highest carbon loss in the top 0-10 cm. Below 10 cm, there was a reduction in the carbon lost in all regions of the meadow and a convergence of values across the dry through wet regions. The intact series of cores showed similar trends, but higher carbon loss in the wet meadow 0-10 cm depth than was found in the composited cores (Figure 3-4).

Averaged across the different water potentials, there was a significant loss of carbon in all of the dry cores, the intermediate 0-10 cm depth, and the wet (0-10 cm and 20-40 cm) depths (Table 3-2). There was a decreasing trend in the total amount of carbon lost in the wet meadow (WT) up to 2.5 bar and then a significant increase at 4.0 bar (Figure 3-5). Although the trend is evident at depth in the wet meadow, only the 20-40 cm depth showed significant differences in carbon mineralization among the different water potentials ($p<0.05$). In this case, the lowest carbon mineralization values were found at the 0.3 and 1.0 bar water potentials, with a subsequent increase at 2.5 and 4.0 bar. This general trend of the lowest carbon mineralization in the 2.5 bar followed by a significant increase was also found in the dry and intermediate regions of the meadow. The intermediate region had the highest values of carbon mineralization at the 0-10 cm depth in the 4.0 bar treatment. The effect was attenuated at depth and no significant differences among water potentials was found. In the dry region of the meadow, there were significant differences apparent at all depths (0-10 cm, 10-20 cm and 20-30 cm). The dry 0-10 cm and 10-20 cm depths displayed the highest flux in the 0.1 bar treatment followed by decreasing carbon mineralization until 2.5 bar. There was a significant increase from 2.5 to 4 bar above 20 cm, but no significant difference below 20 cm. A comparison of the total carbon loss from the composited and intact cores indicates a close alignment of values, with the intact cores showing slightly higher carbon loss than the composited cores (Figure 3-6).

After fitting each replicate with the one pool model (average fitted model shown in figure 4), the initial labile carbon pool was largest in the 0.1 and 4.0 bar treatments in the 0-20 cm depths of the dry and wet meadow soils (Figure 3-7), indicating the presence of two labile pools of carbon in these soils. The intermediate meadow region showed a less pronounced trend in the top layer, but the largest amount of C was mineralized in the 2.5 to 4.0 bar treatments. Below 10 cm in the intermediate meadow, the labile carbon pool was largest in the 2.5 and 4.0 bar treatments. The fraction of the labile carbon pool to the total carbon pool in the meadow was similar above 20 cm in the dry meadow, with labile carbon making up to 15-20% of the total amount of carbon in the 0.1 and 4.0 bar treatments. The wet meadow showed a similar trend although the 10-20 cm depth showed the largest proportion of labile to total carbon. The fraction of labile carbon in the intermediate region of the meadow was highest below the 10 cm depth in the soil in the 2.5 and 4.0 bar treatments. The fitted decomposition rate constant (k) showed the highest rate of decomposition at the 0.3 bar treatment across all moisture regions and depths.

3.5 Discussion

The results clearly show that in most regions of the meadow, the largest carbon mineralization occurs when the meadow is very wet (0.1 bar). The large C loss when meadows are wet is contrary to our original hypothesis that was built on the assumption that, for natural systems that are at or near saturation for long periods of time, drying and increased proportion of air filled pore-spaces would lead to increased decomposition. Hence, we expected that carbon mineralization would peak with drying in the meadow until a critical level of dryness was reached (Baldock and Skjemstad 2000, Van Huissteden et al. 2006). The fact that the largest carbon mineralization occurs in the 0.1 bar treatment suggests that the large pores in the soil were partially drained by this point, allowing air entry into the soil and aerobic decomposition to occur. At this water potential of 0.1 bar, there is still enough water to facilitate diffusion of microorganisms (Carson et al. 2010). As drying progresses (from 0.3 to 1 and 2.5 bar), however, the diffusion of microorganisms is expected to be limited as does the pool of labile carbon (Skopp et al. 1990).

We also observed an increase in C mineralization at the driest water potential (4.0 bar). This phenomenon was also observed in a study by Kechavarzi et al. (2010) that showed increased carbon mineralization with drying. The cumulative carbon mineralization at 4.0 bar was comparable to that of the 0.1 bar treatment in most cases. The correlation between the intact and composited cores was shifted slightly above the 1:1 line indicating that the intact cores had more cumulative carbon loss than the composited cores. This is once again an indication that in these types of soils, structure plays an important role in carbon mineralization (Figure 3-6), as we would normally expect the disturbed (composited) cores to show higher carbon losses.

The initial labile pool of carbon (C_o) of the 4 bar treatment is as large as the 0.1 bar pool. This separation of two pools of labile carbon implies that there is a moisture threshold in these soils that must be crossed before the second pool of labile carbon can be exploited by microorganisms. The notion of having two pools (protected and unprotected) of labile carbon is not new, but has mostly been researched in aggregated grassland and/or cultivated soils (Beare et al. 1994, Mikha and Rice 2004) and not applied to organic soils.

Numerous previous studies have shown that water potential has a significant effect on respiration of microorganisms (Griffin 1981, Skopp et al. 1990, Wilson and Griffin 1975). For example, Wilson and Griffin (1975) found that although bacterial respiration still occurred at dry water potentials (8 to 30 bars), bacteria are highly sensitive to relatively small changes in water potential due to the limited diffusion capability of microorganisms with changes in water content or potential. To demonstrate how hydrology and soil physical properties affect decomposition of soil organic matter, Adu and Oades have also shown that in soils the largest percentage of carbon mineralization occurs in the larger pores as compared to the smaller pores (Adu and Oades 1978). They cite the inaccessibility of smaller pores to microorganisms as a controlling factor on carbon mineralization. The relationship between the volume of habitable pore space for bacteria in the soil and water content of the soil is well established except in soils which experience shrink/swell phenomenon (Postma and Van Veen 1990), where the transient nature of pore volume complicates interpretation of the data. While all studies above conclude that moisture is a constraint on microbial mineralization of soil organic matter, very few investigate the spatial distribution of water that is held in a drying soil and how that relates to the ability of microorganisms to access stored carbon.

There is a typical seasonal cycle of wetting and drying that occurs high elevation wetlands soils, as the main source of water is from a transient snowpack that melts in early spring. Within the historic range of the system, from the water table maximum (WT max) to the water table minimum (WT min), the soils find a tenuous balance between primary production and decomposition, and organic matter accumulates over time (Figure 3-8) (Clymo R. C. 1984). This is due in part to the melting snowpack that recharges the meadow soils creating saturated conditions that prevent significant aerobic decomposition from occurring (Greenwood 1961). Lack of significant summer water inputs to the meadows cause the water table to recede allowing air entry into the pore spaces of the soil. This process of saturation and initial draining can help explain the pattern of carbon mineralization that we find at the wettest water potentials (0.1 – 1 bar), but doesn’t explain how or why there would be a secondary peak in carbon mineralization when the soils are much drier.

3.5.1. Conceptual model for decomposition in high elevation wetland soils

In non-shrinking soils, labile organic carbon is rapidly decomposed if it is not protected from decomposition either within aggregates or in pores too small for microbes to enter (Tisdall and Carter 1996). As soils begin to dry, microbes become limited in their capacity to diffuse and become substrate limited. This process is entirely different in shrinking organic soils where labile carbon is protected not via aggregation, but by the saturation and subsequent oxygen limitation of the soil. In non-shrinking soils, as the soil begins to drydown, the carbon efflux would peak and then decline as water gradually drains from the various soil pores. Since microorganisms are limited by the size of the pore necks and by diffusion constraints, much of this carbon would remain inaccessible for decomposition (Long and Or 2005). We propose that the process is distinctly different in a shrinking organic soil, where larger pores are selectively lost via consolidation, but the microbial-accessible smaller pores remain saturated at a wide range of water potentials until a threshold is reached. After the threshold is breached, these pores begin to drain and a second pool of labile carbon is accessed.

Our conceptual model for decomposition in high elevation wetland soils is divided into six stages (Figure 3-9). The initial stage (WT max) appears when the soils are saturated (directly after snowmelt recharge has occurred), and CO_2 fluxes are low due to lack of aeration in the soil pores. In stage 2, the water table has begun to recede allowing air entry into the soil. This drop in water table causes the largest pores to rapidly drain water and corresponds to an increase in the carbon efflux from the soil as the labile carbon is rapidly mineralized. There is still enough moisture to facilitate diffusion of microorganisms in search of readily mineralizable substrate. Stage 3 corresponds to a further lowering of the water table and a decrease in the total carbon efflux as the water is fully drained from the larger pores and the soil has begun to consolidate as the capillary forces draw the soil particles closer together. This process causes the smaller pores to remain saturated and not drain as they would in a non-shrinking soil. Water becomes limiting for diffusion in the larger pore spaces, restricting the accessibility of the labile carbon to microorganisms. In Stages 4 and 5, the soil is now irreversibly consolidating as the soil is drying past the historic maximum dryness. This drying reduces the carbon flux to lowest levels as microorganisms rapidly consume available carbon before becoming substrate limited. In the final stage, the soil has reached the point where the tensile strength of the soil exceeds the strength of the capillary forces due to drying. At this point, the soil stiffens and the consolidation process ceases. It is only at this point that water begins to release from the finer pores as seen in the water retention curves (Figure 3-2). As a result, a secondary pool of labile carbon becomes accessible to microorganism for decomposition.

The conceptual model presented here becomes even more plausible when we consider fundamental physical properties of soils and how they control soil organic matter decomposition. Soil structure and texture have long been known to be an important control on decomposition of soil organic matter (Bosatta and Ågren 1997, Hassink 1994, Schimel et al. 1994, Van Veen and Kuikman 1990). However, the coupled changes of soil moisture in response to consolidation by capillary suction in wetland soils and the impact this has on carbon mineralization has not been adequately studied in the context of soil organic matter dynamics. However, from geotechnical engineering studies, we know that in non-compressible soils, the internal pore pressures that arise from overburden (or a load) are carried by the intergranular contacts between particles, whereas in compressible soils, this occurs via excess pore water pressure (Croney and Coleman 1954). If that excess pore water pressure is reduced, in a compressible soil, the matrix begins to compact or consolidate.

Even though there are numerous studies that investigated the role of changing hydrology in soil organic matter dynamics, there are only a handful that considered the effects of consolidation. One of the few studies in this area showed that compaction does not have a significant impact on carbon mineralization in the soil (Van Veen and Kuikman 1990). However, detailed reading of the Van Veen and Kuikman (1990) study shows that the compaction tests were performed by applying an external load to the soil which reduces total pore space and forces water from pores of different sizes. It is likely that their results and subsequent interpretations were at least partially affected by the drastic reduction in overall porosity in addition to drying that their samples experienced. Contrary to the method employed by Van Veen and Kuikman (1990), consolidation of soils by capillary suction applies the same basic principles, but leads to more realistic and ecologically relevant representation of how drying induces capillary forces that draw the soil matrix closer together (Arnold et al in prep). Consolidation of soils by capillary suction causes a selective reduction of the larger pore size class, but also results in very small loss in moisture with drying as water is redistributed in the shrinking soil. This is evident in the water retention curves from our study sites.

This study highlights the limitation of our knowledge on the coupled biogeochemical and hydrological interactions of organic soils. We have shown the presence of hydrological thresholds in these soils that can enhance carbon mineralization from these ecosystems in response to drying. This has enormous implications not only for these high elevation ecosystems, but for high latitude ecosystems as well, in that we may see an initial increase in carbon mineralization, but that may be reduced until a critical threshold of drying is reached.

3.6 Conclusions

The results of this study clearly show the presence of a secondary pool of labile carbon that exists in these organic soils. This pool only becomes accessible to the microbial community when the soil undergoes a steady drying event that induces consolidation to the limit of capillary suction. In most cases this secondary pool was as large as or larger than the pool that is decomposed initially upon draining. With an increase in drought-like extreme conditions in the Western United States, there is potential for this pool to be mineralized in the high elevation wetlands. This would lead to widespread meadow degradation and a loss of ecosystem services that these ecosystems provide in the context of mountain hydrology. Furthermore, this results warrants further look into the coupled hydrology and biogeochemistry of wet, organic-rich soils in high altitude and high latitude ecosystems that are expected to experience drying due to anticipated changes in climate.

3.7 Figures

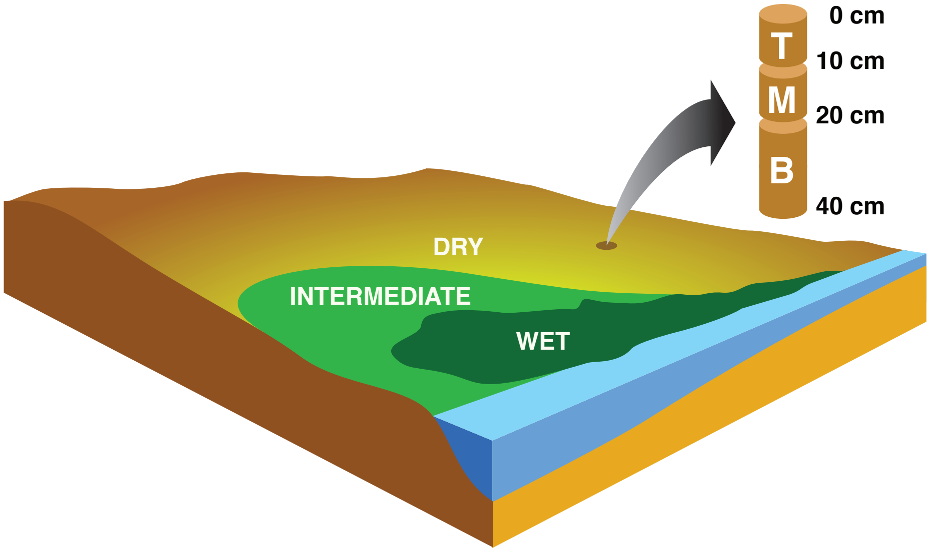

Figure 3-1 Bulk soils and intact cores were collected from three main hydrologic regions in the meadow (dry, intermediate and wet) at three different depths. The sample identifications referred to in the remainder of the manuscript are illustrated in the figure, with D, I, or W, referencing the dry, intermediate or wet site and the T, M, or B referencing the (0-10cm), (10-20 cm) or (20-30 cm –dry only and 20-40 cm) depths.

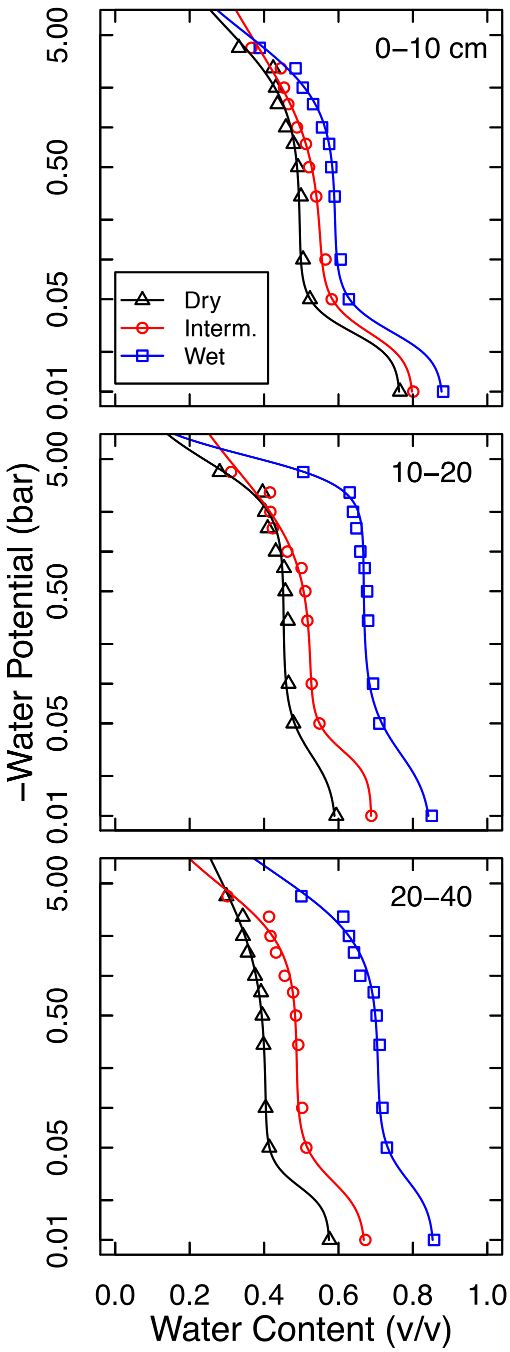

Figure 3-2 Water retention curves for the intact soils cores fitted with the Durner/ vanGenuchtan multimodal retention function (Equation 1).

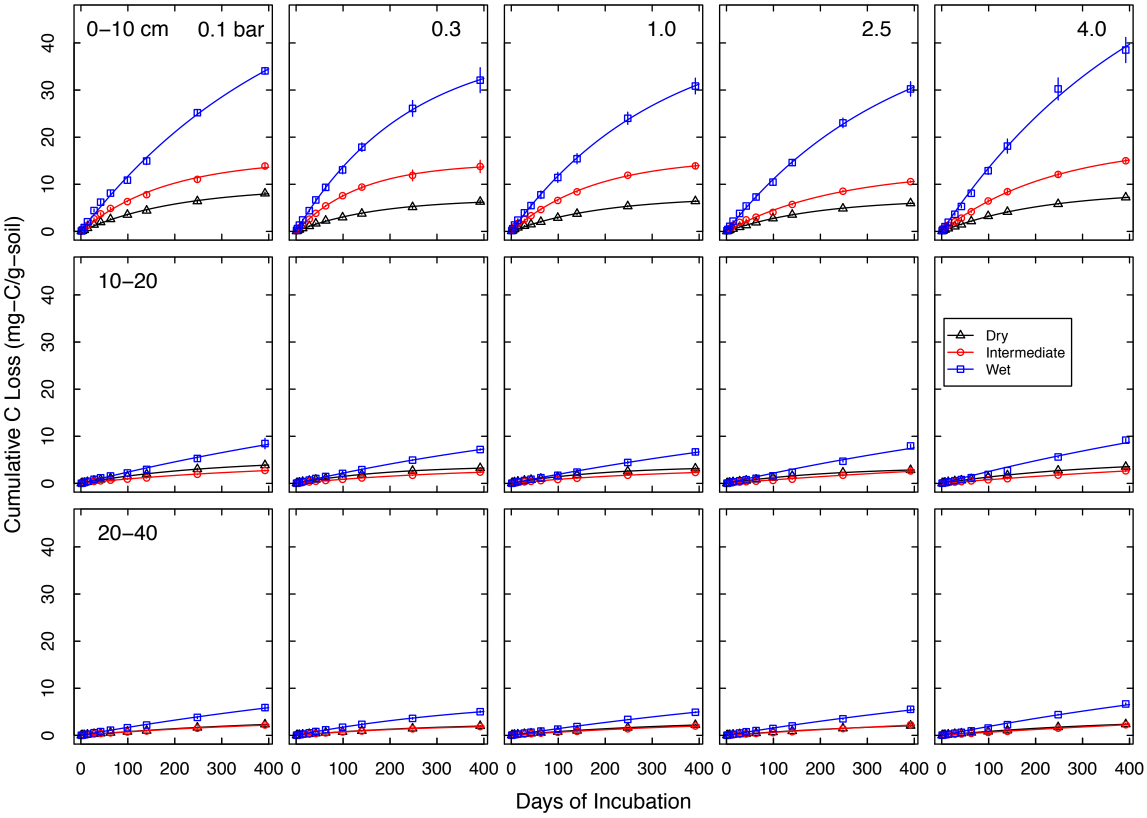

Figure 3-3 Mean cumulative CO2-C evolution data in g CO2-C/g soil for the intact cores (0.3) with the error bars representing standard error. The one pool model (Equation 2) was fitted to the average of the three replicates.

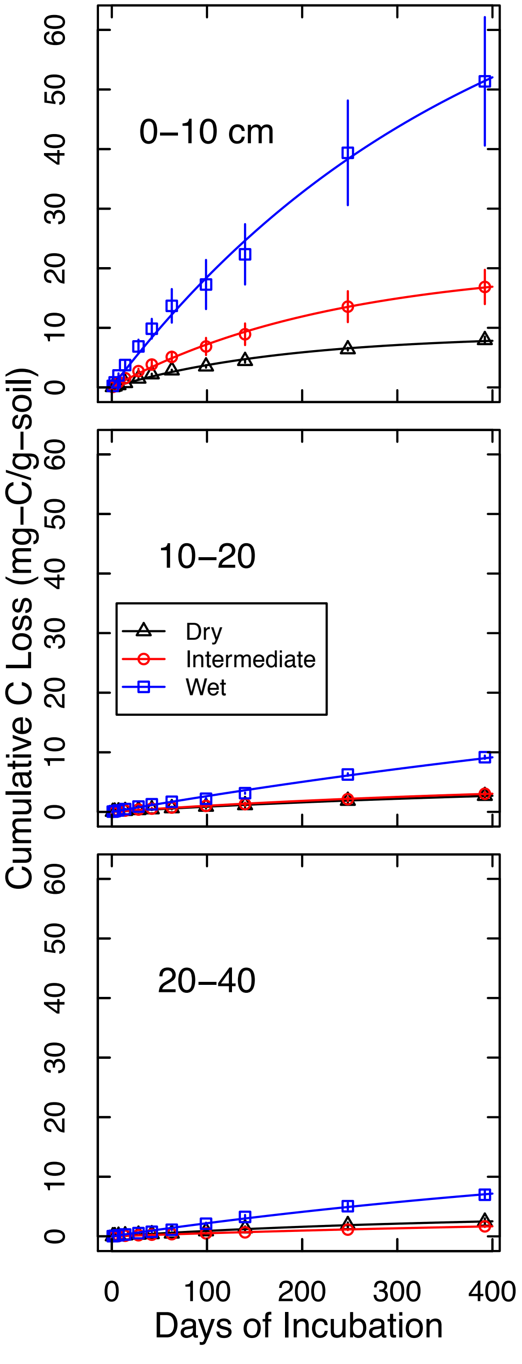

Figure 3-4 Mean cumulative CO2-C evolution data in g CO2-C/g soil for the composited cores with the error bars representing standard error of three replicates. The one pool model (Equation 2) was fitted to the average of the three replicates.

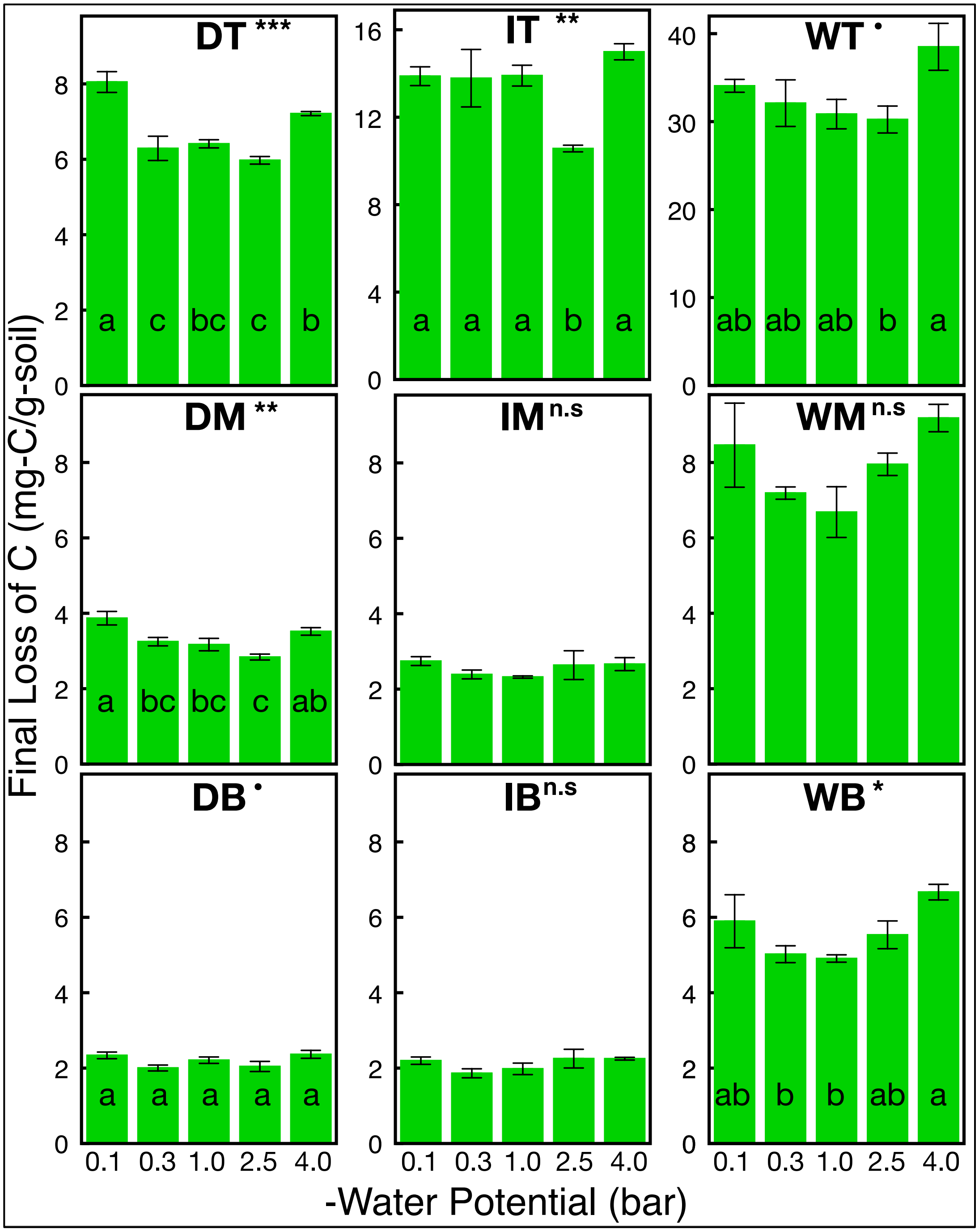

Figure 3-5 The final carbon loss (g CO2-C/ g soil) from the composited cores with error bars representing standard error of three replicates. Lower case letters represent significance determined via a Tukey post hoc test (p<0.1).

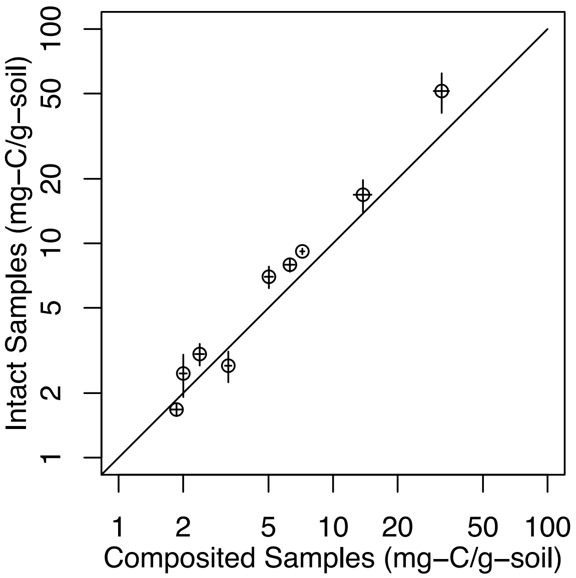

Figure 3-6 The comparison of total carbon loss from composited versus intact soil cores, with the black line indicates the 1:1 line between the samples. The vertical error bars represent the standard error (n=3) of the intact soil cores while the horizontal error bars represent the standard error (n=3) of the composited cores.

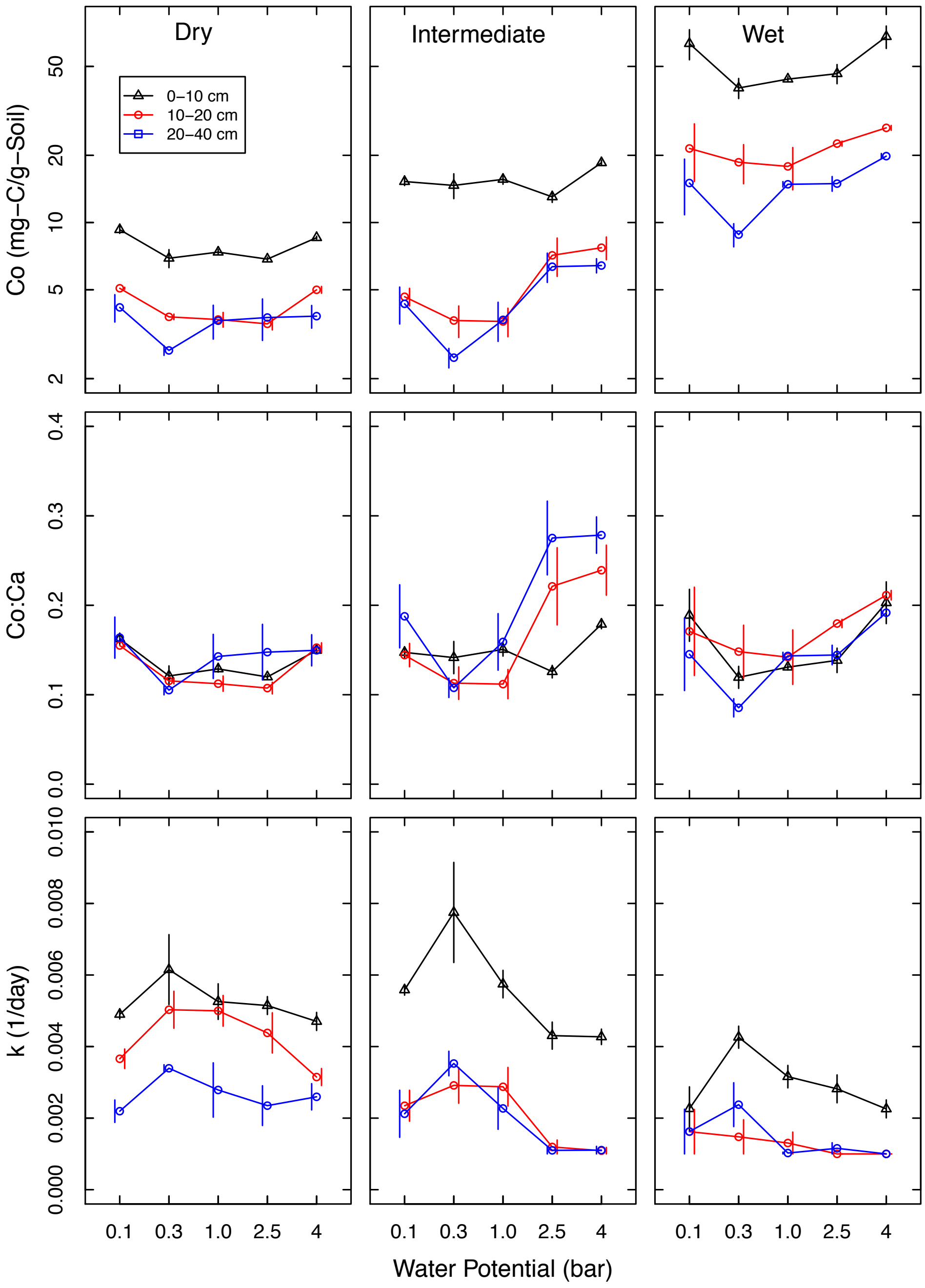

Figure 3-7 Best fit model parameter (C_o, C_o:C_a, and k), where C_o represents the pool of labile carbon, C_o:C_a represents the ratio of the initial labile pool to the total carbon pool (C_a), and k represents the decomposition constant (day^{-1}). Error bars represent the standard error of 3 replicates.

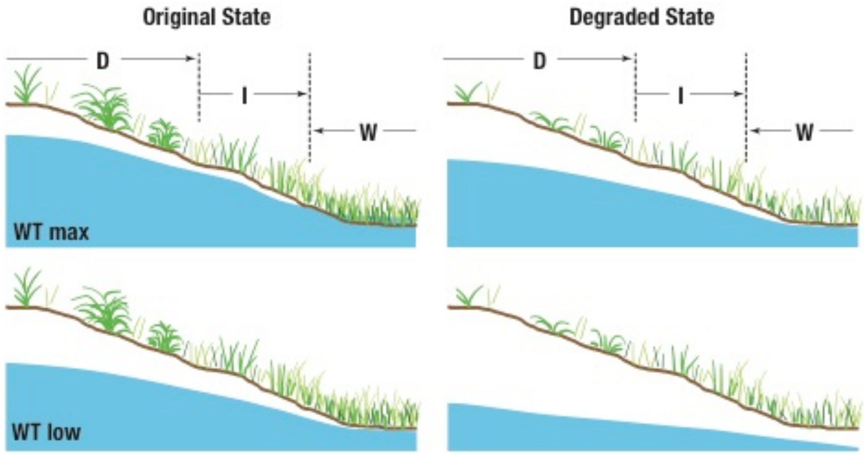

Figure 3-8 Conceptual model for water table range in pristine and degraded states. The dry, intermediate and wet meadow regions are shown with the corresponding water table maximum and water table minimum levels in a pristine and degraded state. In the degraded state, the water table maximum is much lower than in the pristine state, providing conditions for drying of the surface soil layers.

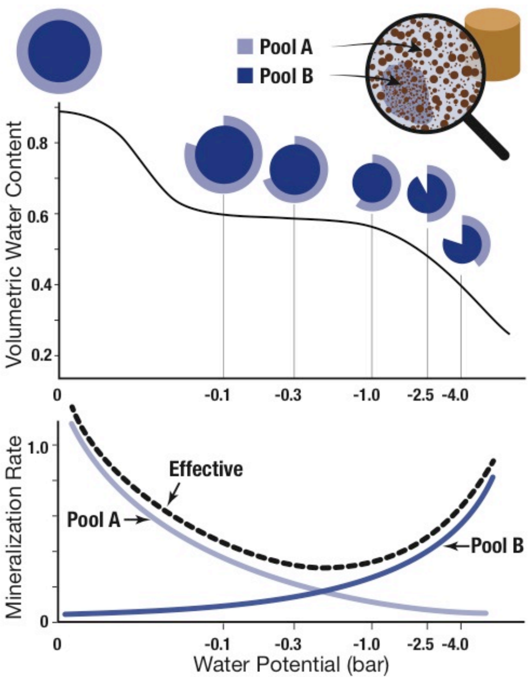

Figure 3-9 Conceptual model of decomposition in high elevation wetland soil. The top panel depicts the actual water retention curve from the wet meadow region. One pool of labile carbon (pool A) is depicted as the light blue outer circle and the second pool of labile carbon (pool B) is depicted as the dark blue inner circle. The lower panel of the figure depicts the effective mineralization rate with drying in the soil, highlighting the presence of two distinct pools of labile carbon.

Table 3-1 Soil properties for the Hall RNA soils are depicted in the table. Standard error is reported for pH and bulk density measurements. Particle size distribution was not completed on the wet meadow 0-10 cm layer due to peat layer.

	Depth	**% Sand**	**% Silt**	**% Clay**	**Total % C**	**Total % N**	**C:N**	**pH (H$_2$0)**	**Bulk density**
Dry	0-10	67	28	5	5.7±0.08	0.39±0.01	14.8±0.02	3.8±0.42	0.89±0.17
	10-20	67	27	6	3.3±0.04	0.24±0.01	13.9±0.15	4.2±0.35	1.26±0.11
	20-30	65	26	9	2.5±0.04	0.19±0.0	13.4±0.18	4.5±0.32	1.56±0.14
Intermediate	0-10	71	23	6	10.4±0.12	0.7±0.01	14.8±0.04	4.5±0.06	0.51±0.06
	10-20	64	31	5	3.2±0.02	0.24±0.01	13.7±0.35	5.2±0.03	1.04±0.06
	20-40	61	32	7	2.3±0.01	0.17±0.0	13.6±0.03	5.4±0.07	1.25±0.03
Wet	0-10	*	*	*	33.5±0.07	2.2±0.01	15.3±0.06	5.3±0.13	0.17±0.01
	10-20	64	32	4	12.6±0.05	0.89±0.01	14.1±0.12	5.7±0.03	0.46±0.07
	20-40	73	25	5	10.3±0.08	0.66±0.0	15.7±0.11	4.8±0.62	0.54±0.16

* not tested

Table 3-2 One-way ANOVA results of final C loss. The degrees of freedom of the treatments (water potential) and the replicates are 4 and 10, respectively.

		Moisture Regime		
		Dry	Intermediate	Wet
Depth	Top	$F = 17.3$, $P < 0.0002$	$F = 6.1$, $P < 0.01$	$F = 2.8$, $P < 0.09$
	Middle	$F = 8.5$, $P < 0.003$	$F = 0.8$, $P = 0.533$	$F = 2.5$, $P =0.11$
	Bottom	$F = 2.7$, $P < 0.091$	$F = 1.4$, $P = 0.295$	$F = 3.5$, $P < 0.05$

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CHAPTER 4. A method for characterizing desiccation induced consolidation and permeability loss of soft soils

Abstract

A new method was developed to measure soil consolidation by capillary suction in organic soils. This method differs from previous methods of measuring soil consolidation in that no external load is utilized and only the forces generated via capillary suction consolidate the soil matrix. This limits the degree of consolidation that can occur, but gives a more realistic perspective on the response of soft organic soils to desiccation in the field. This new method combines the principles behind a traditional triaxial cell (for measurements of volume change), a pressure plate apparatus, (to facilitate drainage by capillary suction), and the permeameter, (to measure saturated hydraulic conductivity), and allows for simultaneous desaturation of the soil while monitoring desiccation induced volume change in the soil. This method also enables simultaneous detection of the historic limit of dryness in the soil, which has important implications for these soils as they respond to a changing climate.

4.1 Introduction

Organic soils are important in the global carbon cycle for their ability to sequester large amounts of carbon but are also highly prized as agricultural land once they have been drained (Armentano and Menges 1986), making these soils particularly vulnerable to anthropogenic modifications. These soils form in regions where the subsurface remains at a condition of continuous saturation to near saturation, for most of the year (Mitsch and Gosselink 2007). Organic soft soils are typically dominated by hydric vegetation that produces an abundance of belowground biomass (Collins et al. 2001). The consistent saturation of the soil in these systems leads to reduced rates of aerobic decomposition and hence accumulation of organic matter in soils with low bulk density (i.e. soft soils) (Collins et al. 2001, Kayranli et al. 2010). Desiccation of organic soft soils either through changes of climate and/or extreme weather events or by draining of land for agriculture can induce both physical and chemical changes in the soil (Preston et al. 1987, Wosten et al. 1997). Consolidation of the soil matrix as a result of capillary forces due to drying can result in widespread land subsidence and a dramatic loss of permeability. Increased rates of decomposition after drying further exacerbates the subsidence over time (Price 2003, Stephens et al. 1984).

From an ecological perspective, to our knowledge, there is no method available to directly measure the degree of consolidation a soil will experience if it is desiccated. This is partly because most soils experience significant wetting and drying cycles over seasonal or annual timescales. Typically in soils, the ability of capillary forces to consolidate the soil matrix is countered by the increasing tensile strength of the soil (Alonso Pérez de Agreda et al. 1990, Baumgartl and Koeck 2004). If a soil has previously experienced drying, the capillary forces generated as the soil dries will not counteract the inherent tensile strength of the soil to induce consolidation. However, this is not the case in soils that have never experienced drying. Previous methods for examining consolidation are derived from the field of geotechnical engineering and have in the past predominantly focused on saturated soils which are consolidated via an external loading (Biot 1941, Terzaghi 1996). In these methods, soils are saturated and then subjected to an increasing external load, which consolidates the soil matrix as water is being pressed out of the pore space. The soil matrix never becomes unsaturated. In 1979, D.G. Fredlund proposed a different method of examining consolidation in unsaturated soils (Fredlund Delwyn G and Hasan 1979), and developed a consolidation apparatus for testing unsaturated soils (Rahardjo 1996). In this method, an external load is still applied to the unsaturated soil to induce volume change. While both of these methods are applicable to the realm of geotechnical engineering, neither are effective at detecting consolidation in organic soft soils that occurs due to increased capillary suction as desiccation proceeds under natural field conditions.

In order to understand how these organic or soft soils will respond to future desiccating events, we designed a new method of measuring consolidation by capillary suction that involves direct measurement of volume change in the absence of an external load. This method expands on previous applications of consolidation of unsaturated soils by combining the principles behind the triaxial cell, (for direct measurements of volume change), the pressure plate apparatus, (to allow for controlled drainage of the soil sample), and the permeameter, (to allow for measurements of hydraulic conductivity). This method also allows for detection of the historic maximum dryness that a particular soil has experienced.

4.2 Theory

4.2.1. Basic concepts

Consider a pair of soft aggregates that are barely touching (Figure 1). As these aggregates are subjected to drying, the capillary water exerts a force that pulls the aggregates towards one another (Ghezzehei and Or 2000, Or and Ghezzehei 2002). Initially, this force is concentrated at a small contact region between the aggregates and the resulting stress (ratio of force to contact area) can be substantially higher than the inherent strength of the soil. In this case, the soil yields to the capillary forces and undergoes deformation. This deformation continues until the contact area grows and the resulting stress no longer exceeds the inherent strength. This process involves both elastic and plastic deformations (Ghezzehei and Or, 2001). The elastic component occurs instantaneously and is swiftly recoverable (Gallipoli et al. 2003) after the deforming force is lifted. The plastic component, however, involves permanent re-arrangement of soil constituents and is usually a time-dependent (viscous) process. Therefore, the deformation stays even after the deforming forces are removed.

An important consequence of this type of deformation is that the soil retains a memory of the maximum stress it has exhibited. When the pair of aggregates shown in Figure 1 are subjected to a second drying cycle, the effective stress felt at the contact will remain below the inherent strength until the capillary forces (dryness) exceeds what was reached in the first cycle. Therefore, when a soil that has experienced prior drying event is subjected to desiccation it exhibits only limited and fully recoverable (elastic) deformation. However, if the historic maximum drying is exceeded, the deformation becomes permanent and irreversible (Alonso Pérez de Agreda et al. 1990).

This process is routinely observed in freshly prepared seedbed made up of a loose assemblage of fine aggregates. Such aggregate beds rapidly get compacted as they are subjected to repeated wetting and drying cycles. The fact that capillary forces dominate this process was shown experimentally (Silva, 1995 cited in Ghezzehei) and theoretically (Ghezzehei and OR, 2000). Similar concepts are also routinely employed in geotechnical engineering

(Dunn et al. 1980). The stress induced by structures is usually engineered so that the stress imparted by the foundation does not exceed the historical maximum overburden (commonly a referred to as “pre-compression stress”) (Junior and Pierce 1995). Upon rewetting a slight rebound in volume occurs if the deformation was only elastic. Therefore, the relationship between suction and porosity can be considered as analogous to the relation between external loading and porosity, generated using geotechnical oedometer tests (Figure 2). One of the main differences in this method is the desaturation of the soil sample results in a concomitant increase in the tensile strength of the soil matrix with drying. Ghezzehei and Or showed that soil strength increases exponentially with soil drying using rotational rheometer (Ghezzehei and Or 2000). In Figure 3, utilizing a subset of the data we illustrate that the yield strength (stress at which plastic deformation is initiated) and viscosity (rate of plastic deformation) are both influenced by soil wetness. The data shows that the tendency for both capillary and plastic deformation increases by orders of magnitude as a soil dries. This non-linear effect of soil rheological properties severely limits the maximum consolidation that can occur via capillary suction. Another difference with this method is that the stresses that are induced via desiccation originate within the soil and act in all directions simultaneously unlike stresses that are applied externally, which are inherently unidirectional and must act from the boundary.

The ability to detect the pre-consolidation stress in soils is a critical step in foundation engineering and geotechnical engineers have devised methods that accurately determines this important characteristic of soils. The most accurate approach involves subjecting a soil specimen to sequentially increasing stresses. The measurement, which typically is conducted using saturated soils, permits the soil to reach equilibrium deformation at stress step. Finally, the stress sample volume (or porosity) data are plotted on a log-linear graph. It has been shown that slope of the stress-volume relation undergoes a drastic shift at the pre-consolidation stress (Casagrande 1936). This method is fundamentally important in compressible soils because it allows for separation of the elastic, reversible consolidation and the plastic, irreversible consolidation and allows detection of the the prior stress history of the soil (Holtz and Kovacs 1981). The strength and plasticity of a soil is influenced by the past stress history in the soil, which in turn is determined by the amount of organic matter, water content and texture of the soil as well (Nawaz et al. 2013). The amount of organic matter in a soil has also been shown to influence the degree of compressibility of a soil, with an increase in organic matter coupled to a decrease in compressibility (Soane 1990). However, Smith et al. (1997) found that coupled soil texture and organic carbon content were key to assessing forest soils most at risk for compressibility (Smith et al. 1997).

4.2.2. Effect on flow properties

It is well known that the structure of a soil has profound influence on the ability of water to move through the soil profile (Hillel 1980). The size distribution of pores in the soil dictates the water retention capacity as well as the permeability of the soil (Dullien 1992, Fredlund DG et al. 1994). In organic soft soils, the degree of decomposition is also related to the pore size distribution and hence permeability with less decomposed soils having a higher hydraulic conductivity (Boelter 1969). The process of consolidation by capillary suction initially functions to deform and compress the macropores in a soil, which has an immediate and dramatic effect on the permeability of a soil by reducing the ability of water to preferentially move through the matrix (Or et al. 2000, Price and Schlotzhauer 1999). This also alters the naturally occurring structurally heterogenous state of most soils to a more homogenous pore distribution (Horn et al. 1995). This forces water through alternative pathways (meso and micro pores) in the soil, resulting in a reduction of permeability even for a small change in volume of the soil sample.

The goal of this study is to develop a method that allows detection of the pre-consolidation stress (historical maximum dryness) by analyzing capillary suction versus soil volume data. From an ecological perspective, this allows for determination of how past climate has influenced desiccation in the soils that have developed in humid regions with minimum past incidences of extreme desiccation. This method will apply to soils that have not been subjected to severe drying which would induce a change in strength of the soil and render the limit of dryness undetectable.

4.3 Materials and Methods

4.3.1. Study site and soil sampling

The research site is located in the Harvey Monroe Hall Research Natural Area, located along the crest of the Sierra Nevada, California. Intact soil cores were collected in a high subalpine meadow that formed in glacial till. The particular meadow is hydrologically connected to an alpine lake via an outflow stream as well as a seepage face through the glacial till. The soils in this region are young, having transitioned to peat accumulation around 4500 years before present. There are two known mega droughts that occurred during the formation of these soils (Stine 1994). However, the soil that would have been impacted by desiccation during these droughts is located just above and below a dated tephra layer from the Inyo-Mono crater chain (Wood

1977). The tephra layer is located between 15 and 20 cm deep in the meadow soil. The intact cores that were utilized for the development of this method were taken from the top 0-5 cm depth, to eliminate possible historic impacts of drought. The soil cores were gently removed from just below the top layer of moss using a soil hammer equipped with a soil bucket and plastic liner. Care was taken that the samples were not compressed during the process. Intact cores were stored on ice and transported to the laboratory where they were kept at 5° C.

4.3.2. Consolidation equipment design

The consolidation experimental setup consists of the utilization of three separate methods (triaxial cell, pressure plate apparatus, and permeameter), which allows for simultaneous draining of the sample with direct monitoring of volume change of the sample. In addition, the system can reverse flow and resaturate the sample, enabling the determination of saturated hydraulic conductivity at various points during the experiment.

The apparatus consists of a modified trixial cell (Geocomp Corp., Acton, MA). The base of the triaxial cell was fitted with a custom designed base containing ceramic plate (Figure 4). To enable saturated hydraulic conductivity measurements, we used high-permeability ceramic disc with bubbling pressure of 0.5 bar (Soilmoisture Corp., Santa Barbara, CA). Airtight tubes fitted below the base of the porous disk were connected to a high-precision piston pump (Geocomp Corp., Acton, MA) that permits direct control of volumetric flow rate or fluid pressure. This pump controls the flow of water into and out of the soil sample, allowing for saturation and subsequent desaturation of the sample. The soil sample sits on the porous disk and is encased in a latex membrane (GeoComp), and fitted with O-ring seals at the top cap and base. The latex-membrane encased sample is placed in a thick plexiglass cylinder that is sealed at the top and bottom with O-rings and bolts. Airtight tubes run from the cap at the top of the soil sample out via connections fitted at the bottom of the plexiglass cylinder. The plexiglass cylinder surrounding the soil is filled with water and connected via airtight valves and tubing to a water filled bladder (Part # HM 4151A, Humboldt Mfg. Co., Elgin, IL). The water pressure is controlled by an automatic pressure controller unit (Part # HM 4151A, Humboldt Mfg. Co., Elgin, IL). The volume of water that flows in to the triaxial cylinder to substitute for soil sample volume loss is accurately determined using a flow-through volume change apparatus (Part # HM-2315 Humboldt Mfg. Co., Elgin, IL) fitted with a digital LVDT (Part # HM-2310.10 Humboldt Mfg. Co., Elgin, IL). An airtight tubing connected to the automatic pressure controller unit supplies compressed air to the top end of the soil sample. The automatic pressure controller unit is internally fitted with a differential pressure regulator such the pressure of the water jacket is always at 5 KPa. This pressure difference is sufficient to keep the latex membrane in contact with the soil but not strong enough to cause substantial soil deformation.

Data from the automatic pressure controller unit (confining water pressure and air pressure at the sample top), the LVDT (change in sample volume) and the piston pump (water pressure at the bottom of the sample and the volume of water delivered to or removed from the soil) are continuously streamed to computer and logged by software supplied by the respective manufacturers of the devices.

4.3.3. Experiment Design

4.3.3.1. Sample installation

An installation cylinder (GeoComp) is utilized to stretch the latex membrane and install the soil core on the porous disk. The plexiglass cylinder is gradually filled with water via the by utilizing the bypass feature on the volume change apparatus and care is taken to remove any entrapped air bubbles in the cylinder. Once the cylinder is full of water, the system is pressurized in small steps including increasing the pore pressure below the porous disk and increasing the air pressure to the top of the sample to both equal 100 KPa. The water jacket is pressurized to 105 KPa and the pressure regulator is set to maintain that constant 5 KPa effective pressure regardless of external temperature and pressure fluctuations in the laboratory. Once the system is pressurized and stable, the pore pressure is increased by (0.5 to 1 KPa) to saturate the soil sample from below. The sample is allowed to saturate until equilibrium is reached and water is flowing out of the top of the sample. At this stage, a hydraulic conductivity test can be conducted.

4.3.3.2. Hydraulic Conductivity Test

A hydraulic conductivity test is performed on a saturated sample by subjecting the soil sample to a gradient of pore-water pressure (ΔP), with the bottom of the sample having higher pressure. Water is allowed to flow upwards through the sample until equilibrium volumetric flow rate (Q) is achieved. This steady flow process can be described by Darcy's law as;

$$Q = K\frac{A}{\rho g L}\Delta P \qquad (1)$$

where,

K = hydraulic conductivity (cm/s)

A = cross sectional area of sample (cm^2)

L = Length of the sample (cm)

P = Pressure (KPa)

ρ = density of water (g/ml)

g = acceleration due to gravity (cm/s^2)

Because, all the quantities in the equation except K are independently known, the saturated hydraulic conductivity can be calculated directly. To provide a better constraint on the calculated hydraulic conductivity and to ensure that the assumption of linearity that is implied in Darcy's law is obeyed, we conducted the experiment at four different levels of pressure gradient. Thus, when we plot the volume gradient versus the pressure gradient the data must fall on a straight line. The slope of the line then uniquely identifies the hydraulic conductivity of the soil.

4.3.3.3. Capillary Consolidation Test

To begin a consolidation test, the sample is initially saturated and allowed to equilibrate for several days. The pressure is then lowered in the pore pressure pump in small increments (~5 KPa) to the desired pressure level and the sample is allowed to drain and equilibrate over several days to weeks. At each pressure step, the effective pressure of 5 KPa is maintained by the pressure regulator between the top of the sample and the water in the plexiglass cylinder. As the sample desaturates, volume changes induced by the increase in capillary suction in the soil sample triggers the volume change apparatus to move more water into the plexiglass cylinder to maintain the 5 KPa effective pressure. The amount of water that is injected is then equivalent to the degree of deformation of the soil sample due to the desaturation. The test proceeds with the amount of water removed from the sample as well as the change in volume being monitored through time at each pressure step. This allows for careful detection of the pre-consolidation stress which can then be equated to the historic maximum dryness the sample has experienced.

4.4 Results and Discussion

4.4.1. Experiment A – Determination of Pre-consolidation stress - Consolidation curve

In order to determine the preconsolidation stress, an intact core collected from the seepage face of the study site in the summer of 2012 was utilized. The sample is a fibrous peat taken from directly below the moss layer (0-5 cm). The total initial volume of the sample was 83.095 ml. The sample was installed on the porous disk and the plexiglass cylinder filled with water. The system was pressurized in small steps (10 KPa) to 149 KPa and the sample was allowed to equilibrate overnight. In the first step, the pore pressure pump was lowered by 5 KPa and the sample allowed to equilibrate for several days. At each successive step, the pressure was lowered 5 KPa, until a final pressure of 54 KPa was reached.

As the first step was initiated, there was a large change in the volume of water that was released from the sample in comparison to the magnitude of volume change in the sample (Figure 5). This indicates that the sample is desaturating as well as consolidating. The pressure step was not lowered until the volume of water extracted leveled off and equilibrium was reached.

Upon completion of the testing protocol, the sample volume change data was converted to a relative volume by normalizing it against the initial sample volume. The relative volume was plotted against the capillary suction (water potential) intervals (Figure 6). The consolidation curve shows a distinct change in slope that is indicative that the experiment exceeded the pre-consolidation stress that this sample had seen in the past. The pre-consolidation stress point is indicative of the historic maximum dryness that this particular soil has experienced in the field. First we visually determined which data points to fit with a line and then fit each set of the log linear data points with a straight line. A separate equation describes each line and because they pass through a single point, that point can be determined algebraically. It is this point at the intersection between the two lines that reflects the historic maximum dryness point. With a historic maxima of 0.163 bar, we can conclude that this soil has formed in an environment conducive to a high water table, which allowed for an accumulation of organic matter. This allows for better understanding of conditions of soil formation as well as the ability to predict the response of the soil to future desiccating events. The fact that the soil's historic maxima is much wetter than most soils illustrates the future sensitivity of these types of soils to drying.

4.4.2. Experiment B - Hydraulic conductivity determination

Another important aspect of this method is the ability to determine saturated hydraulic conductivity at different points in the experiment. This can allow for tracking of the loss of permeability in a sample as a direct result of the consolidation process. In this experiment, another intact core from the same location and depth as the sample utilized in Experiment A was installed on the porous disk via the same method as above. The system was pressurized the same as Experiment A, and the sample was allowed to saturate overnight. At four separate pressures, we measured volume flow in ml over time. At a given pressure, the slope of the line represents the volumetric flow rate (Q) in ml/s (Figure 7a). After measuring Q at multiple pressure steps, we plot the measured Q values versus the pressure steps and fit the points with a straight line (Figure 7b). The slope of the fitted line is then equivalent to the saturated hydraulic conductivity value for the soil.

At this point, an initial hydraulic conductivity measurement was made on the sample by increasing the pore pressure by steps of 1 KPa. The initial value of hydraulic conductivity for the sample was 0.0003 cm/s. Upon completion of the initial hydraulic conductivity test, the sample was drained in 10 KPa steps until 50 KPa suction was reached and then the sample was allowed to equilibrate at 50 KPa of applied suction. This process took approximately 11 days. The flow in the pore pressure pump was reversed and the sample was allowed to re-saturate over several days. The irreversible loss in volume of the soil sample can be detected as the change in the volume from the initial saturation to the volume at re-saturation (Figure 8). The volumetric flow rate was calculated. When the sample was in equilibrium, a second hydraulic conductivity test was made. In this case the total sample volume lost was 2.47 ml however there was a 64% loss of hydraulic conductivity at 50 KPa (Figure 9). The sample was resaturated again and the test repeated with 75 KPa suction. Results show that the sample experienced consolidation with an additional volume loss of 1.91 ml, which was smaller than the initial consolidation at 50 KPa. The final hydraulic conductivity test showed an overall 76% loss in total hydraulic conductivity.

4.5 Challenges and considerations

This method has indicated its sensitivity to detecting the historic limit of dryness in soft compressible soil samples and it is this sensitivity that renders the system susceptible to environmental fluctuations of temperature and pressure (laboratory conditions). We mitigated these challenges in the following ways:

1. This experiment requires simultaneous and accurate regulation and recording of air pressure (at the top of the sample) and water pressure (at the bottom of the sample and the confining vessel). Because water and air respond to temperature changes, this experiment is extremely sensitive to temperature fluctuations. A drop in temperature results in more significant contraction of compressed air than compressed water. This results in an apparent increase in the confining pressure and suction experience by the soil. Part of this could be real physical change, which will be sensed by the devices and results in corrective feedback. In general, this type of change results in noisy data. It may also make detection of equilibrium difficult, which requires long duration of equilibration. Earlier designs of our experiments used a piston pump to control the confining pressure, but this turned out to be very sensitive to fluctuations. We partly mitigated this challenge by using a single automatic pressure control unit that supplies regulated air pressure through a pair of outlets with a fixed differential pressure. Then, the confining pressure is applied by pressurizing a bladder with the highest outlet from the pair of the outlets.

2. Due to fact that viscosity (the rate of deformation) increases exponentially as soils dry (see Figure 3), the time needed to reach equilibrium increases with each time step. It is important to wait long-enough (several weeks) at the higher suction levels in order to make sure that the deformation reached completion.

4.6 Summary and Conclusion

The method for detecting consolidation by capillary suction in the absence of an external load offers the ability to determine the effect of future desiccating events on organic soft soils that have not been subjected to extreme drying. While organic soft soils are estimated to cover 4-6% of the global land area, they account for more than 20-25% of the organic carbon stored in the soil (Mitra et al. 2005). The unique ecosystems that have allowed for development of organic soft soils provide an abundance of ecosystem services including storage and filtering of water, flood attenuation, nutrient removal and habitat, all of which are fundamentally linked to the physical and chemical properties of these soils. While the effects of draining on these soils has been studied, the influence of natural climate variability and the consequences of desiccating events on the ability of these soils to perform their ecological roles has not been examined. This method allows for an analysis of the historic maximum dryness that the soil has experienced, and also allows for predictions on how the soil will respond to future desiccation.

This method is significant in that it advances our ability to understand and quantify the soil structural response to a changing climate. With this method, we now have the ability to detect memory of past historical maximum dryness experienced by given soil. This is very important as we study the sensitivity of soils in susceptible environments that may go to environmental conditions that have not been experienced in recent past. In addition, it allows projection of what degree of change in terms of sample volume and sample permeability can be expected under projected climate change scenarios.

4.7 Figures

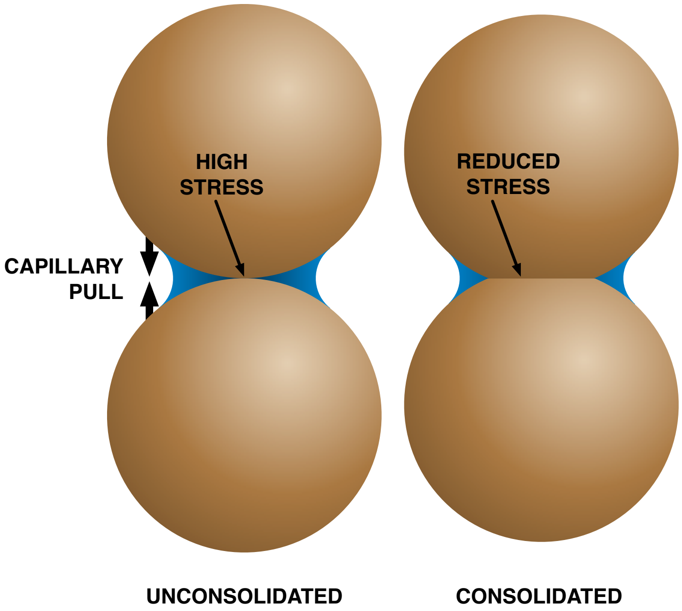

Figure 4-1 Unconsolidated and consolidated aggregates subjected to a wetting and drying cycle that induces the deformation of the unconsolidated aggregates through by capillary pull.

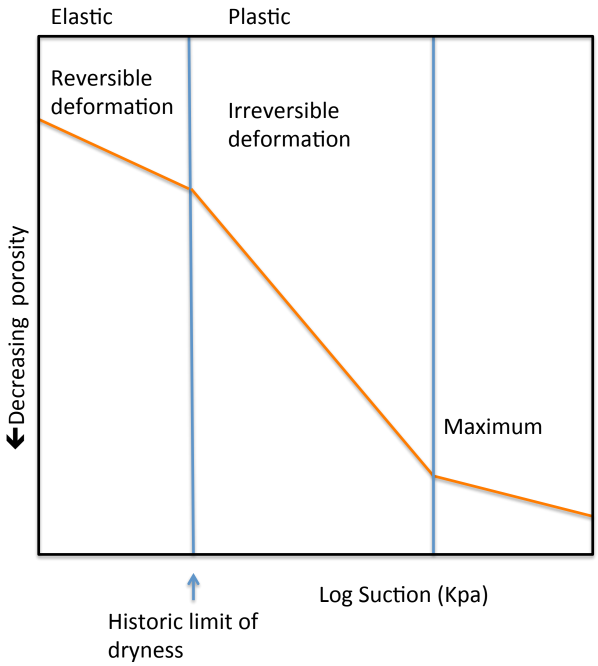

Figure 4-2 Theoretical soil consolidation curve showing the reversible (elastic) deformation and the (plastic) deformation. The historic limit of dryness is determined by the intersection of the elastic and plastic deformation lines.

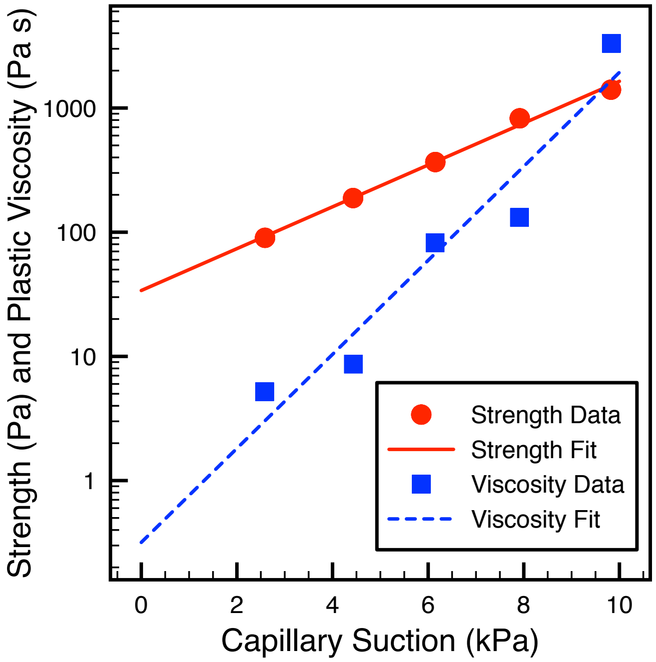

Figure 4-3 Rheological data from Ghezzehei and Or (2000) highlighting existence of increase in strength and plastic viscosity of a Millville silt loam soil with dryness.

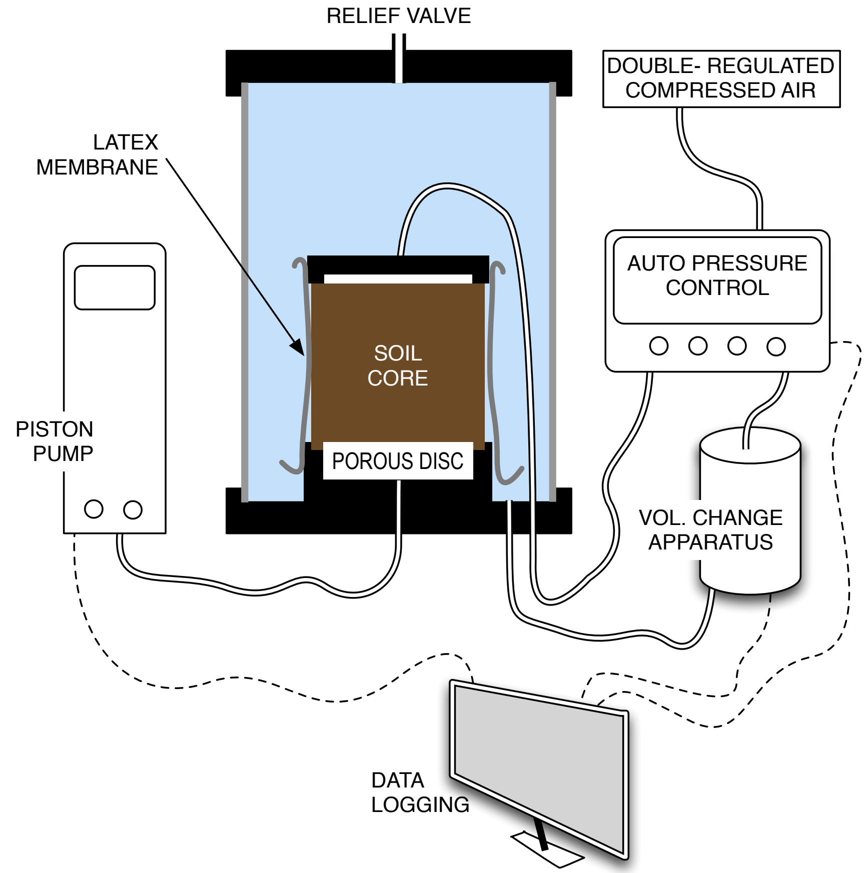

Figure 4-4 Schematic of Experimental Setup

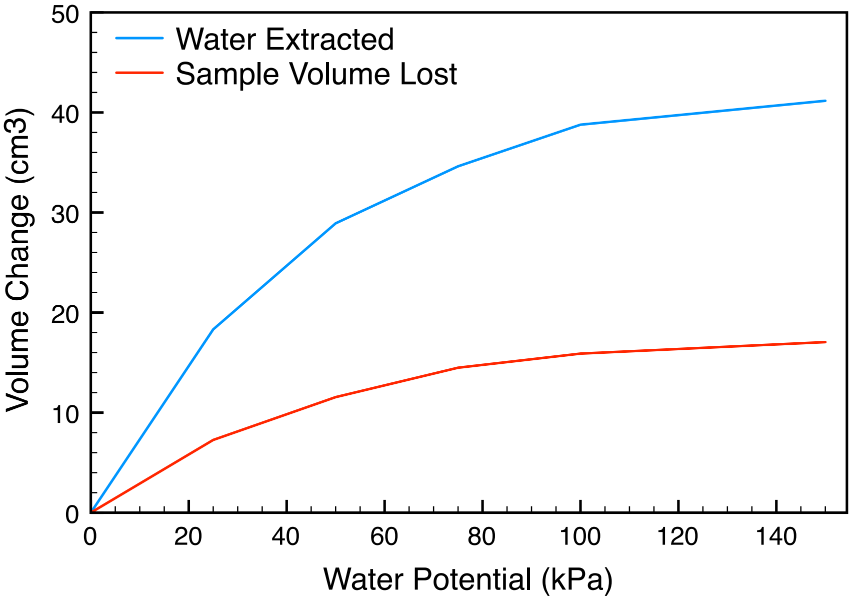

Figure 4-5 Example of soil sample and water volume changes with increasing suction.

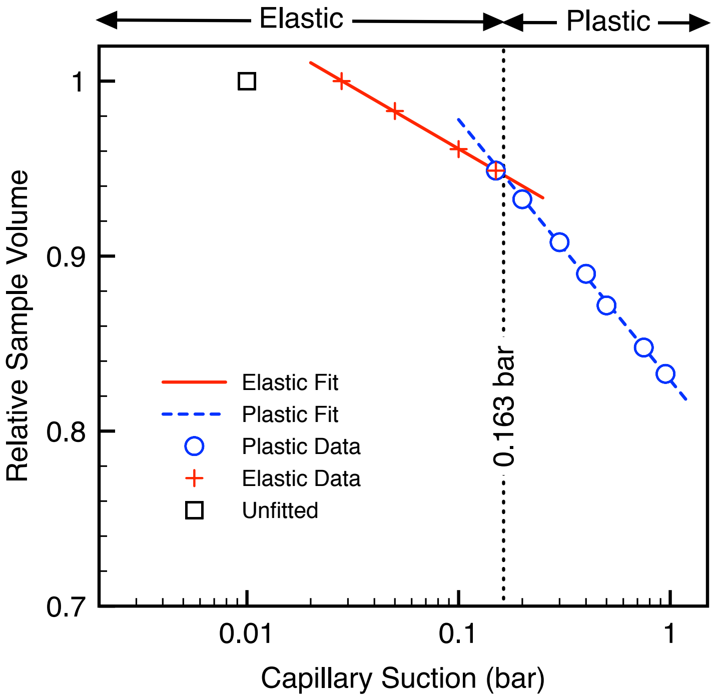

Figure 4-6 Consolidation curve based on an intact soil core collected in 2012. The red line is fitted to data points in the elastic region of the curve, and the blue line is fitted to data points in the plastic region of the curve. The black symbols represent data that was not fitted. The intersection between the two lines represents the historic limit of dryness.

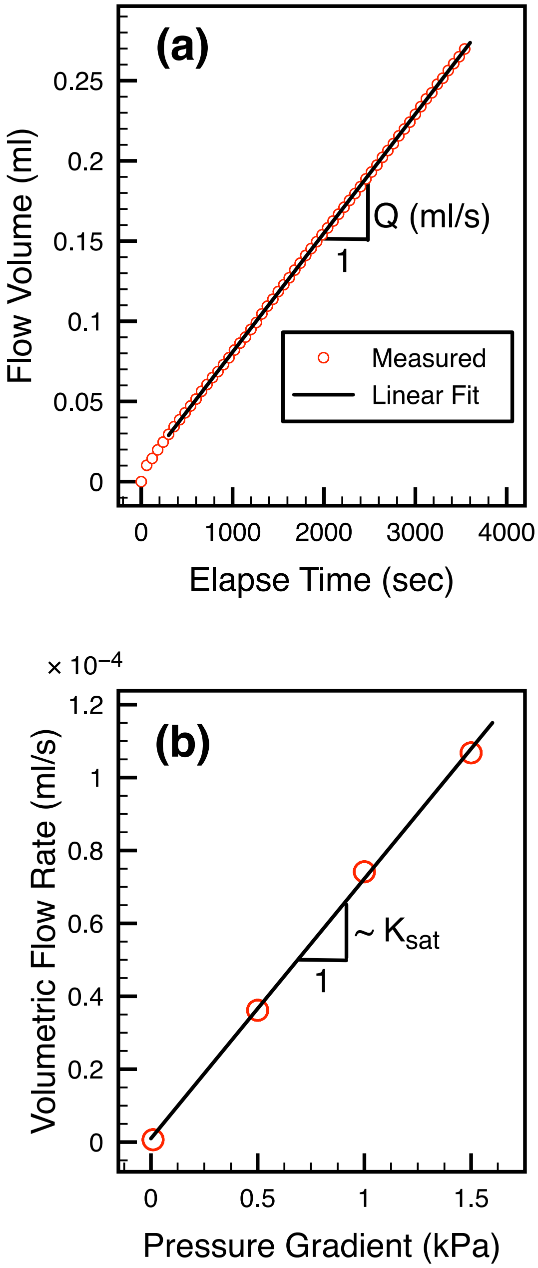

Figure 4-7 Determination of Hydraulic Conductivity with sample pressure step volumetric flow rate determination (a). The value of Q from each pressure step is plotted in panel b to determine the saturated hydraulic conductivity (Ksat).

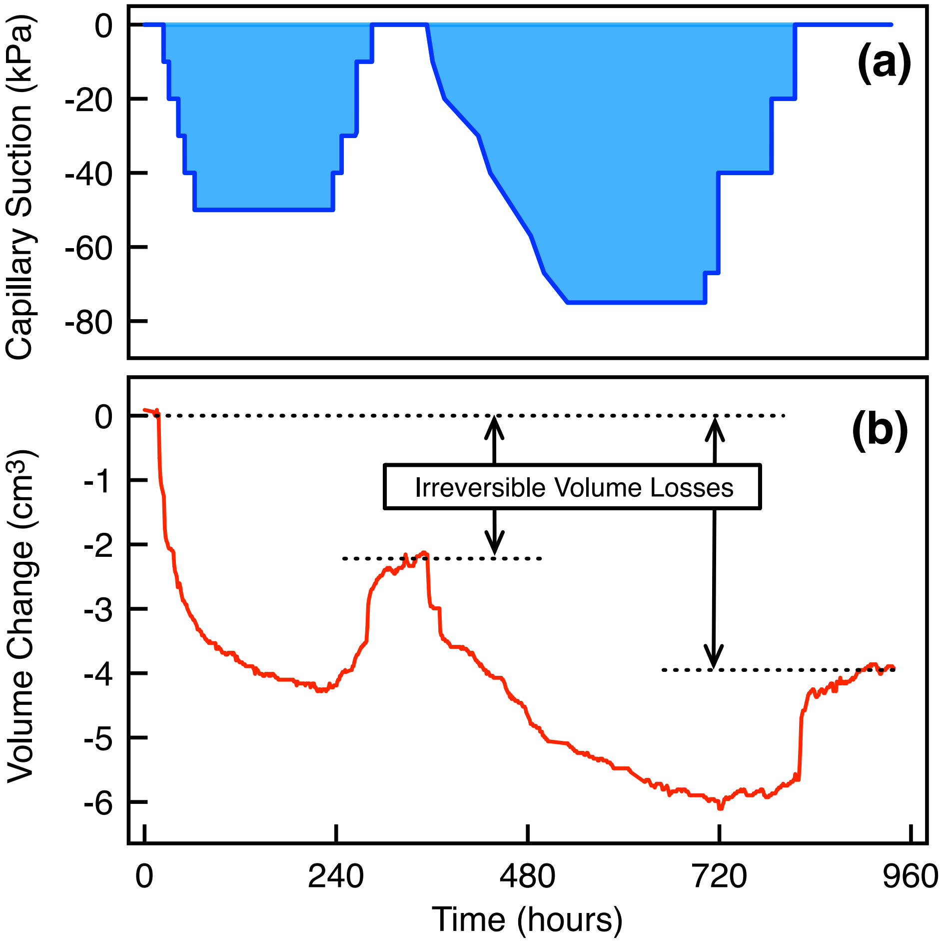

Figure 4-8 Determination of irreversible volume change in the soil sample. Panel a) depicts the water removed from the sample with each test, and panel b) illustrates the subsequent volume change measured.

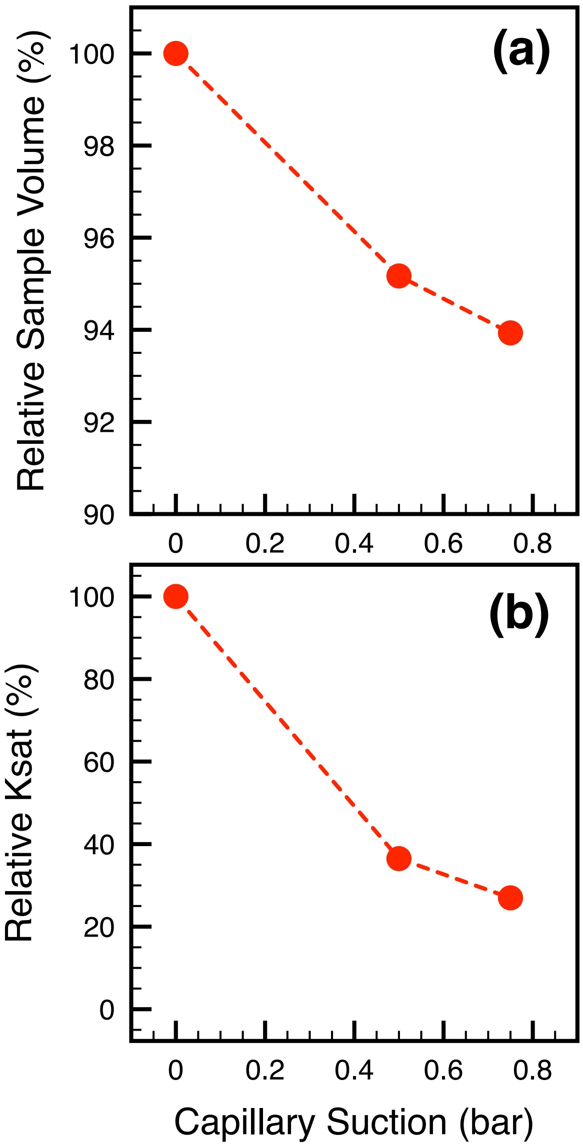

Figure 4-9 Changes in sample volume (a) and Ksat (b) in Experiment B at two levels of drying (0.5 bar and 0.75 bar).

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CHAPTER 5. Exceedance of historic limit of dryness triggers hydrological tipping point in high elevation peatlands

Abstract

While snowpack provides the dominant water storage reservoir in mountain regions, the soil plays a fundamental role in secondary water storage and is a critical and often overlooked component of mountain hydrology. With their large quantity of stored carbon, high elevation peatland soils provide essential ecosystem services related to water storage, filtration and slow release to downstream communities; and carbon sequestration. Extreme dry years can cause the water table in these systems to drop below historic levels. This drying can induce changes in the structure of the soil through capillary consolidation, coupled to a simultaneous change in the rate of decomposition of soil organic matter. Using a multiple method approach, we investigate the historic limit of dryness that high elevation peatland soils in the Central Sierra Nevada have experienced in order to determine if future drying can trigger a hydrological tipping point resulting in an irreversible loss of ecosystem services. We found that within the historic limit of dryness (up to 0.04 bar suction), high elevation peatlands are resilient and accumulating carbon. After exceeding the 0.04 bar dry limit, however, the peatlands begin to consolidate leading to loss of porosity and permeability, and loss of soil carbon through decomposition. In addition we show that the structural changes in the soil are rapid, have immediate consequences for high elevation peatland resilience and have a disproportionately large impact on hydrology in comparison to decomposition. This research highlights how small changes

in climate can trigger local hydrologic tipping points in mountainous regions with cascading regional scale impacts.

5.1 Introduction

The resilience of any ecosystem hinges on its ability to respond to perturbations while maintaining its normal functionality (Holling 1973). The loss of ecological resiliency renders ecosystems vulnerable to further perturbations, and ecosystems are particularly vulnerable to perturbations as they approach tipping points that can lead to major alterations and/or loss of ecosystem stability and services (Dai et al. 2012). The actual tipping point in any given ecosystem often goes unnoticed until it manifests itself in a visible change, for example as loss of sea ice, biota range shifts, receding glaciers, etc. (Lenton 2013).

Most past research on ecosystem response to changes in climate has focused on effect of increasing atmospheric temperature and warming: including implications of warming for biogeochemical cycles of carbon and nitrogen (Conant et al. 2011, Melillo et al. 2011)and fire (Marlon et al. 2012, Moritz et al. 2012). However, there is now a growing body of scientific literature on ecosystem responses to changes in precipitation patterns (Berhe et al. 2012, Chou et al. 2008, Cruz-Martínez et al. 2009, Daly and Porporato 2005, Fay et al. 2008, Knapp et al. 2002, Suttle et al. 2007) indicating that perturbations to the hydrologic cycle may have much more significant consequences, than warming, on ecosystem resilience (Gleick 1989, O'Gorman and Schneider 2009, Porporato et al. 2004, Trenberth 2011). Globally, ecosystems characterized by long wet (and cold) periods, such as peatlands, permafrost, and wetlands, contain large stocks of carbon (Baldock and Broos 2011, Brady and Weil 2008) where soil carbon is stabilized against microbial decomposition largely due to the hydrology and related ecosystem conditions (Schmidt et al. 2011). These ecosystems contain over 20 % of vegetation C and over 50% of the soil C stocks globally (Jones et al. 2010). Soil carbon stored in these hydrology controlled ecosystems is sensitive to changes in climate (Armentano and Menges 1986) where lifting of the hydrologic limit on microbial activity is expected to speed up the rate of OM loss with subsequent implications for reduced plant productivity and altered soil physical and chemical conditions (Chapter 2 and 3). A question still remains as to how much disturbance can hydrology-dominated ecosystems withstand before reaching a tipping point and undergoing irreversible change in ecosystem stability.

Snow dominated mountain ranges, such as the Sierra Nevada Mountains, are the source of water for most of the communities of the western United States. These ecosystems rely on high elevation peatland meadows to filter, store and slowly release water to the river systems (Hammersmark et al. 2008, Loheide II 2009). Over the last decade, we observed changes to the depth and duration of snowpack in addition to an increasing rain-snow elevation line (Johnson et al. 1999, Pupacko 1993) in the Sierra Nevada of California. However, there is still uncertainty as to the direction and magnitude of the response of the high elevation peatlands to these changes in precipitation patterns. It is clear that a dry year in mountain regions can have significant economic consequences that reach far beyond the downstream communities (Medellín-Azuara et al. 2008), through a reduction of water allocation to agricultural, industrial and residential users. However, it is not fully understood what the long-term impact of one or a few dry years will be on ecosystem services delivered by these mountain ecosystems.

In addition to their essential roles on filtering, storing, and slow release of water, the high elevation wet meadows of the Sierra Nevada also play critical roles in mountain hydrology through their ability to attenuate floods (Viers 2013). These peatland systems are dependent on a high water table to slow rates of decomposition and allow for accumulation of carbon (Jungkunst and Fiedler 2007). The accumulation of carbon in these regions has led to the formation of peat beginning around 4500 years before present (Anderson and Smith 1994). The water table range (from water table maximum to water table minimum) in these systems is intrinsically linked to the climate of the region (Eagleson 1978) and forms the predominant control on soil formation. While other soil forming factors are important (topography, biota, time and to a lesser extent, parent material) (Jenny 1941), without a wet hydroclimate to maintain a high water table, these peatlands would not persist (Wood Spencer Hoffman 1975).

There is now a documented shift in the onset of spring (Cayan et al. 2001) and a forecasted intensification of hot extremes in the Western United States (Diffenbaugh and Ashfaq 2010) that has led to an urgent need to better understand how these peatland systems will respond to drought. A significant drop in the water table can cause desiccation-induced consolidation (Berglund and Berglund 2011, Dawson et al. 2010) which leads to a loss in porosity and permeability in the soil. In addition, the lowered water table increases the rate of decomposition and further reduces the water holding capacity of the soil (Kechavarzi et al. 2010). Any given soil can be characterized by a unique limit of dryness that it previously had experienced (due to past and present climatic conditions). In organic-rich, peatland soils, the historic limit of dryness exerts control on present soil physical and biogeochemical processes and allows for prediction of likely changes into the future. For wetland and peat soils, the dryness limit is at a fairly high water content, as they have a high water table that is maintained by surface and subsurface flow regimes. Organic soils experiencing relatively small change in the range of the water table shrink and swell elastically, with transient changes to the soil hydrologic properties (Schlotzhauer and Price 1999). However, if the water table drops below historic levels, exposing the soil to desiccation, the historic limit of dryness is breached and irreversible consolidation will occur (Figure 1). Irreversible consolidation progresses by selective plastic deformation of the larger pores in the soil (Peng et al. 2007). This process proceeds as the soil continues to experience drying, until the tensile strength of the soil exceeds the capillary forces drawing the matrix together (Guérif 1990). At this point the consolidation process will stop, but can be followed by secondary consolidation due to settling and/or further decomposition (Kennedy and Price 2005, Price and Schlotzhauer 1999).

The ecological implications of consolidation are well known in peatlands as rapid subsidence has been observed as a consequence of draining (Nieuwenhuis and Schokking 1997, Schothorst 1977). Most studies have focused on the impacts of a artificially lowered water table through the draining of peatlands for agriculture, or involve laboratory manipulations of water table in microcosms to examine how draining impacts biogeochemical cycling (Moore and Knowles 1989, Van Huissteden et al. 2006). However, so far, we have incomplete understanding of the degree to which natural climate extremes can influence consolidation of organic soils and the ecosystem services that we derive from them.

The past four years (2011-2014) in California have brought extremes on both the wet and the dry ends of the precipitation spectrum presenting a natural experiment to test the effects of the exceedance of the historic limit of dryness on ecosystem processes. The year 2011 was the seventh wettest year on record based on snow water equivalent records from Dana Meadows, Yosemite National Park, California, that extend back to 1929 (Chapter 2). The subsequent two years, 2012 and 2013 were marked by extreme drought, with 2012 and 2013 ranked in the top five driest years in the historic record (1929-2014). The current year (2014) is 12th. With this extreme change in precipitation, there was a dramatic shift in the onset of spring, which occurred approximately 57 days earlier in 2012 and 2013 (Chapter 2). Our objective in this study was to determine the historic limit of dryness of the high elevation peatland soil in 2011 and utilize that dryness as a zero point to determine the impact of future drying on the hydrology and coupled carbon cycling of these systems.

5.2 Methods

5.2.1. Soils

Intact peat soil cores (85 ml) were collected at the end of the summer in 2011 and 2012 from the top 5 cm of soil (below moss layer) along a seepage face in an upper limit subalpine meadow (3200 m elevation) in the Harvey Monroe Hall Research Natural Area, Inyo National Forest of California. The seepage face is formed from a perched water table in glacial till that is hydrologically connected to an alpine lake. Soil-forming processes in this region began approximately 10,000 years before present, as the last major glaciers receded (Curry 1969). The smaller glacial advances (Recess and Matthes) have been shown to be mainly confined to mountain cirques and did not advance through the study site after this point (Clark and Gillespie 1997). However, although mineral soils were forming, peat accumulation did not occur until approximately 4500 years before present due to a changing hydroclimate (Anderson and Smith 1994). A tephra layer forms a distinct layer in the soil at 15-20 cm depth and was deposited from an eruption that occurred in the Mono-Inyo crater complex around 1190±80 years before present (Wood Spencer H 1977). The two most recent mega-droughts in California occurred before AD 1112 and before AD 1350 (Stine 1994). By taking a sample from the top 0-10 cm of the soil profile, we eliminated the possible legacy effects of past extreme drought on the soil structure (Figure 2).

5.2.2. Experimental Design

5.2.2.1. Soil consolidation

In order to induce consolidation by capillary suction, 2011 and 2012 soil cores were installed on a modified triaxial system (Figure 3) that was designed to simultaneously drain water from the samples while quantifying volume change (Chapter 4) and water content of the soil sample. Complete runs for each sample take two-four months. We measured samples from 0-5cm depth that were collected in 2011 and 2012. After installation, the system was pressurized to maintain an effective pressure of 5 kPa between the air pressure supplied to the top of the sample and the water jacket surrounding the core. This effective pressure maintained contact between the latex membrane and the soil sample and prevented water from flowing through gaps between the membrane and the soil. The soil sample was saturated from below overnight and allowed to equilibrate before the tests were initiated. Hydraulic conductivity was measured before the start of the first test by using the saturated pressure as the zero point and increasing the pressure by 0.05 bar and allowing water to flow upward through the soil (Chapter 4). This allowed us to determine the volume of water that moved through the soil over time at any given pressure step.

At the completion of the initial hydraulic conductivity test, the soil was drained at small increments. The pressure steps were changed only after the sample volume ceased changing. After the last pressure step, the water pressure below the sample was increased to allow water to flow back through the sample slowly until the sample was completely resaturated. A second hydraulic conductivity test was completed after saturation was reached. The historic limit of dryness was determined by utilizing the Casagrande method for determining the pre-consolidation stress point (Casagrande 1936). We followed the same procedure for the 2012 soil core (0-5 cm depth), but continued to drain the sample until 1.0 bar was reached. For the hydraulic conductivity measurements, we saturated a 2012 soil core from 0-5 cm depth and took an initial hydraulic conductivity measurement. The soil core was then drained to 0.5 bar. After resaturation, a second hydraulic conductivity measurement was conducted. We drained the samples one more time to 0.75 bar, resaturated and took a final hydraulic conductivity measurement.

5.2.2.2. Water retention and Carbon mineralization

The following methods for water retention and carbon mineralization were described in detail in Chapter 3, but are briefly summarized here:

For determination of water retention curves, paired soil cores were collected in 2011 and one core of each pair was saturated overnight. Subsequently, these samples were sequentially drained on the pressure plate apparatus to the following suctions (0.01, 0.05, 0.1, 0.15, 2.0, 3.0, 4.0 bar) to generate a water retention curve. At each pressure step the sample was weighed. At the conclusion of the test, the samples were dried at 105°C for 48 hours. The second core of each pair was saturated overnight and drained to 0.3 bar, weighed and incubated for over one year in a one liter mason jar covered with plastic wrap. Water was added to the samples on a weekly to biweekly basis to maintain the initial drained weight of the sample. At the end of the incubation, the cores were resaturated and then drained on the pressure plate to the same suctions as the first pairs in order to generate a final water retention curve.

In order to better understand the impact that soil consolidation has on the mineralization of carbon, we conducted a long-term soil incubation using composited soil samples collected from peat deposits (subset of data from Chapter 3). The three samples from the 0-10 cm depth (n=15) were repacked in cores to native bulk density, saturated from below and drained on a pressure plate apparatus to one of 5 different water potentials (0.1, 0.3, 1, 2.5

and 4 bars). The sample cores were then incubated at 20°C for 392 days. Water was added weekly if needed to maintain the initial drained weight. Gas samples were drawn at intervals over the duration of the incubation and analyzed for carbon dioxide on a Shimadzu GC2014 fitted with a thermal conductivity detector.

5.2.2.3. Surface flux of carbon dioxide

We installed 10 cm diameter soil collars in the peat meadow study site located at 3200 m elevation in the Harvey Monroe Hall Research Natural Area, Inyo National Forest, California. Initially, there were 2 collars in 2011, but we increased the number of collars to 6 in 2012. Surface CO_2 flux was measured weekly to biweekly from the start of the growing season until mid October in 2011 through 2012. Measurements were made with a LICOR 8000A portable infrared gas analyzer (LI-COR Biosciences, Lincoln, Nebraska USA), fitted with a portable 10-cm soil respiration chamber. After a 45-second pre-purge, one-minute measurements were recorded and were followed by a 30-second post-purge (subset of data from Chapter 2).

5.3 Results and Discussion

5.3.1. Determination of historic limit of dryness

A 2011 soil core from the 0-5 cm depth in the peat meadow was drained on the modified triaxial system and the historic limit of dryness was reached at 0.04 bar suction (Figure 4). The sample was further drained to 0.3 bar and then resaturated. The gentle initial slope of the data (up to 0.04 bar) indicates that the consolidation is elastic. This deformation and associated changes in soil permeability would be restored completely when the soil is resaturated. Past 0.04 bar the data follows a steeper slope suggesting that the mode of deformation has transitioned from elastic to plastic. This mode of deformation involves permanent re-arrangement of soil constituents that cannot be restored even after the suction force is lifted (Ghezzehei and Or 2001) Thus, the transition water potential of 0.04 bar can be considered as a signature of the historic maximum stress the soil has experienced.

A similar test was repeated utilizing a 2012 sample from the 0-5 cm depth, which was drained to a final suction of 0.9. The historic limit of dryness was detected at 0.17 bar suction. This indicates the possibility that conditions in 2012 were dry enough to irreversibly consolidate the peat surface soil. The shift in the historic limit of dryness from 0.04 in 2011 to 0.17 bar in 2012 is plausible given historic meteorological data along with research that has shown a limited snowpack, warmer spring and summer temperatures, and winter desiccation of meadow soils (Chapter 2). This combination of limiting moisture coupled to warmer temperatures caused severe desiccation of meadow soils in 2012, with cracking of the surface peat observed in the field (Figure 5). To the best of our knowledge, this research represents the first time a signature of historic dryness in soil has been detected, and represents a significant contribution to furthering our understanding of the soil response to rapid climate change.

Because it is impossible to predict the historic maximum stress prior to the start of the test, it was not possible to stop the drainage at the critical suction level and determine hydraulic conductivity. However, we can measure saturated hydraulic conductivity at several points during the test, and then extrapolate back to estimate the change in saturated hydraulic conductivity that would result from irreversible consolidation in the soil. Figure 6 illustrates the results of hydraulic conductivity tests on the 2011 and 2012 peat samples. Up to the historic limit of dryness in each sample, the saturated hydraulic conductivity is not expected to change. However, once the historic limit of dryness is exceeded, the changes to the saturated hydraulic conductivity become permanent. It is important to reiterate here that these hydraulic conductivity values correspond to *saturated* soils that have been drained to different levels of suction. When the 2011 sample was drained to a final suction of 0.12 bar, we measured a new saturated hydraulic conductivity of 2.14e^{-4} cm/s. The 2012 sample was drained to two levels of suction (0.5 and .75 bar) to test the effects of further drying. We found that the 2012 initial saturated hydraulic conductivity was surprisingly consistent with the 2011 final saturated hydraulic conductivity at 2.0e^{-4} cm/s, giving further evidence of the irreversible change in soil structure that occurred in the summer of 2012 in peat meadow soils. Past the historic limit of dryness in 2012, there is a dramatic reduction in hydraulic conductivity to 7.30e^{-5} at 0.5 bar and 5.39e^{-5} at 0.75 bar.

After exceeding the limit of dryness, we find that the total volume loss in the 2011 sample at 0.12 bar was 7%, which equated to a 44% loss in permeability (Figure 7). For the 2012 sample, we found that the 0-5 cm sample lost over 4.5% volume at 0.5 bar with a resulting 64% permeability loss and 6.0% volume loss and a 73% permeability loss at 0.75 bar suction. The relatively small losses in volume coupled to large losses in permeability highlights the fact that even small amounts of drying past the historic limit of dryness have can significantly alter hydraulic properties of these soils.

5.3.2. Impact of changing soil hydraulic properties on carbon mineralization

In order to determine if the exceedance of the historic limit of dryness in peat meadow soil alters carbon mineralization, we conducted a long-term soil incubation with composited peat soil cores drained to different water potentials (0.1, 0.3, 1, 2.5 and 4 bars). We found that the peat meadow soil loses the most carbon at the highest and lowest water potentials, indicating the presence of two labile pools of carbon in the soil that are only accessible after a hydrologic threshold is breached (Figure 8). The reduction of carbon losses at 1.0 and 2.5 bar show that the effect of exceedance of the historic limit of dryness can be attenuated slightly by the shrinking soil matrix, which serves to reduce large pores sizes, and redistribute water among the remaining pores rather than draining it. An examination of the water retention curves illustrates how little water is drained from the soil after the initial drainage to 0.05 bar (Figure 9). The smaller pores remain saturated until approximately somewhere between 2.5 and 4 bar suction, when the tensile strength of the soil exceeds the capillary forces as a result of drying. This effect of saturation of small pores may be the reason why carbon losses are lower with increasing water potential until the soil stops shrinking. This is in opposition to what would be expected with a non-shrinking soil, where drainage of water from small pores reduces the diffusion capacity of microorganisms and they quickly become substrate limited (Skopp et al. 1990).

The positive relationship of organic matter concentration with overall water holding capacity is very well acknowledged in soil science literature (Baldock and Broos 2011, Brady and Weil 2008, Rawls et al. 2003), but how that may vary with accelerated decomposition under changing climatic conditions is currently unknown. To understand the impact of increased decomposition on water retention, we generated water retention curves (WRC) before and after a long-term incubation of peat cores to determine if consolidation and decomposition impact the water holding capacity of the soil. The paired cores were collected in 2011, and their expected limit of dryness was 0.04 bar. Three paired replicates were analyzed from the 0-5 cm depth in the peat meadow soil. The replicates are remarkably consistent, with replicates 1 and 3 showing a drop in saturated water holding capacity post incubation of 12% (Figure 9). Replicate 2 showed no change in water holding capacity post incubation. The WRC show a bimodal pore distribution, with the largest drop in moisture found from saturation to 0.05 bar. This is consistent with the draining of the large pores in the soil. After the 0.3 bar suction, the paired cores are both consolidating, but only the "after" cores have also decomposed. The "before" cores show a gradual drop in water retention with an increasing water potential, but the “after” cores do not drain water as readily. This may indicate a connection between changes in soil chemistry as a result of decomposition and water holding capacity that warrants further investigation.

In addition to the lab-based experiment, we monitored surface fluxes of carbon dioxide from the first snow free day in the peat meadow until October 15 in 2011 and 2012. The wet, short growing season of 2011 resulted in cumulative carbon losses of 519.53 ± 5.3 g C/m^2/growing season, whereas the very dry and prolonged summer of 2012 doubled the cumulative carbon losses (1122.66±22.2 g C/m^2/growing season) from the peat meadow soil (Figure 10).

5.3.3. Impacts on ecosystem services

In most ecosystems, it is hard to pinpoint an event or circumstance that can cause a tipping point to a regime shift. Usually this is because it isn’t investigated until after a regime shift has occurred. In hydrology dependent ecosystems, the exceedance of the historic limit of dryness can trigger a tipping point that irreversibly changes soil hydraulic properties. Due to the coupled nature of hydrology and carbon cycling in these systems, there are both positive and negative feedbacks that can either maintain ecosystem resilience or trigger a regime shift as a result of these changes. An example of a negative feedback would be how the shrinking soil matrix can also function as a “self preservation” mechanism (Price 2003). As the soil desiccates and consolidates, subsidence occurs, which lowers the level of the soil surface. That serves to keep the soil saturated as it subsides closer to the water table, reducing carbon losses. In these high elevation peatlands however, the soil depth is so shallow, that this feedback would be minor. An example of a positive feedback would be that desiccation of meadow soils causes a selective destruction of large pores and decreases permeability of the soil. This can cause more snowmelt to runoff the surface rather than infiltrate into the soil, leading to further drying of the soil over the growing season, when precipitation is minimal (Figure 11).

Another consequence of consolidation is the impact to the hydric vegetation in the peat meadow. It has been shown that mosses have a difficult time recruiting and establishing in consolidated peat soils (McNeil and Waddington 2003). Restoration of drained peatlands have emphasized surface moisture as key in establishing mosses and other hydric vegetation (Komulainen Veli - Matti et al. 1999, Waddington and Price 2000). Continued drying can subsequently result in vegetation changes and shifts in peat soils (Komulainen Veli-Matti et al. 1998).

The most significant impact of the exceedance of the historic limit of dryness is without a doubt the instant and irreversible change to the ability of meadows to store and release water to the streams. This impact will greatly reduce the ability of meadows to deliver ecosystem services that are critical for downstream communities. While the changes to soil structure are immediate and irreversible, processes that might enhance the permeability such as frost heave or burrowing animals can impact hydrology over the long term. However, any subsequent changes to the soil structure will not bring back the original structural state of the soil but will result in a new soil structure that may or may not improve soil hydrologic properties. The coupled biogeochemical response to the drying enhances decomposition, which will further reduce saturated water holding capacity in the peat soil. It also appears that coupled decomposition and consolidation processes function to retain the water in the soil subjected to severe drying rather than releasing water. The loss of macroporosity in the peat meadow soil reduces the amount of water that will be delivered to the streams over the summer drydown. The impact of consolidation and decomposition on the filtering capacity of the peat meadow soils was not investigated in the study and remains an area of future research.

5.4 Conclusion

In hydrology-dependent ecosystems such as peatlands, the exceedence of the historic limit of dryness can trigger a hydrological tipping point that causes a irreversible loss of the capacity to deliver ecosystem services. Relatively small losses in porosity of the meadow soil were coupled to large changes in hydraulic conductivity. The majority of this loss occurred at the early stages of drying. The future implications of this change in hydrologic properties of the soils in this system are an increase in the proportion of snowmelt to runoff rather than percolation into the soil, and a reduction of ecosystem services that meadows provide for downstream communities. This includes reduced flood attenuation capability, and a reduction in the contribution of meadow water to stream baseflow. In addition, we expect an increase in decomposition with drying which will further reduce the water holding capacity of the meadow soils. This research highlights how small changes in climate can have a much larger and consequential impact to society than previously understood.

Due to the sensitivity of this method to environmental conditions in the laboratory (i.e. changes in temperature and pressure) and the time consuming nature of the method, we have analyzed only a few cores thus far. Our integrated approach has clearly shown that the signature of extreme dryness leaves detectible change in soil properties (evidence of tipping-points). There is a need to expand and conduct this kind of measurements to establish a long-term trend in soil changes due to climate change and determine the ecosystem impacts.

5.5 Figures

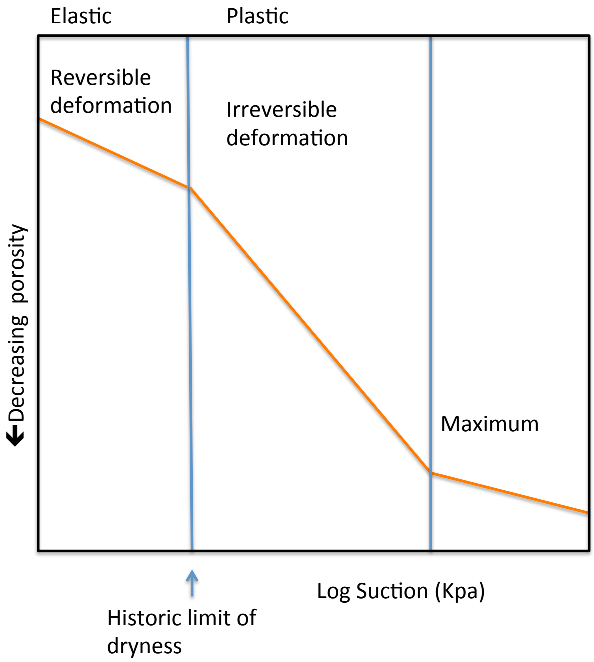

Figure 5-1 Typical consolidation curve for an organic soil showing the elastic and plastic deformation that occurs with drying (reprinted from Chapter 4).

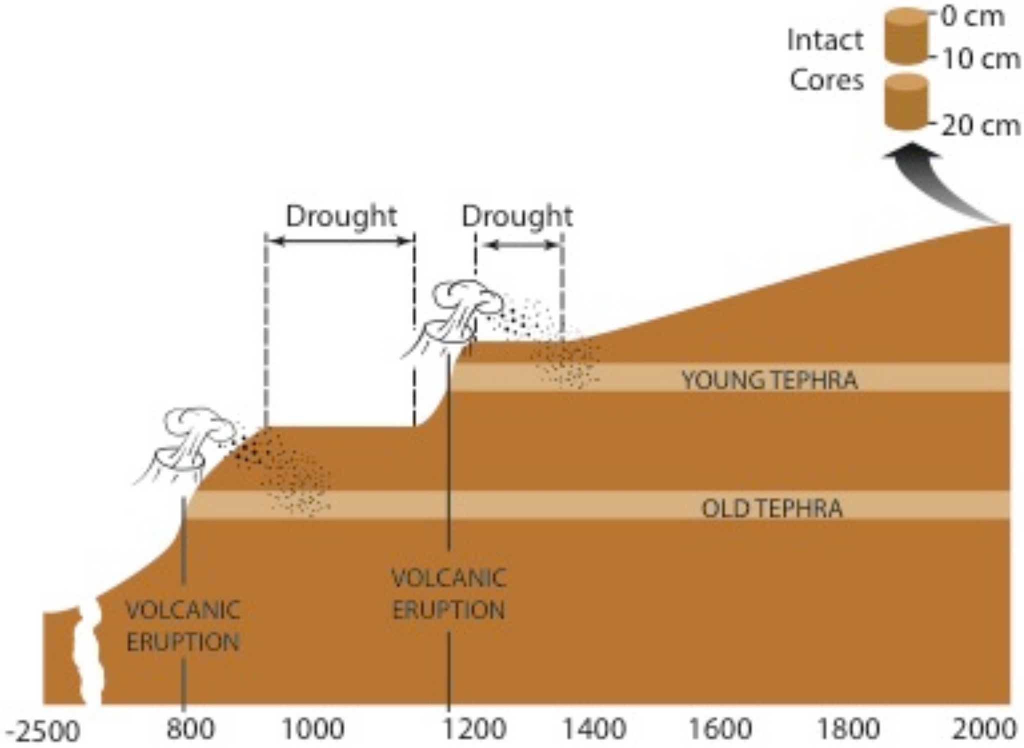

Figure 5-2 Soil formation diagram showing the presence of the tephra layers in the meadow soil and approximate timeframe of the last two mega-droughts.

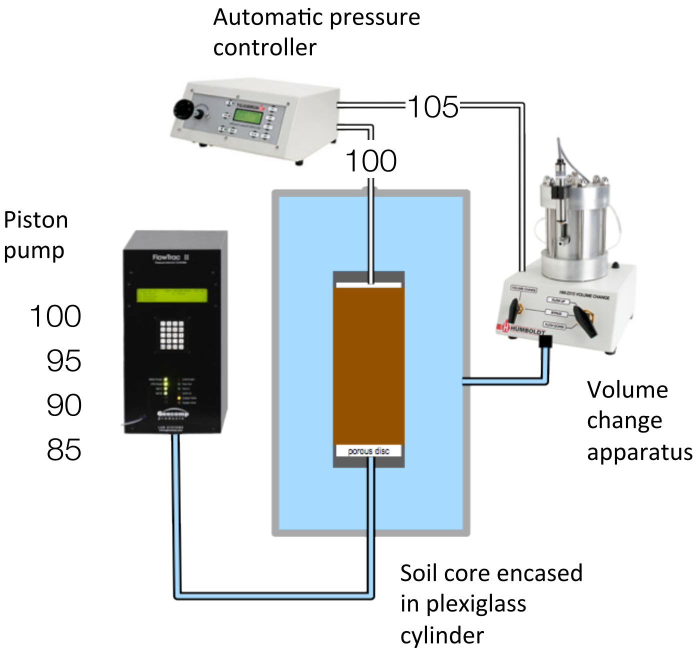

Figure 5-3 Modified triaxial system to measure desiccation induced consolidation by capillary suction (Chapter 4). Number values from the automatic pressure controller indicate the pressure of the air to the top of the sample (100 KPa) and the pressure to the water inside plexiglass (105 KPa). The piston pump values represent the pressure steps the sample was drained at.

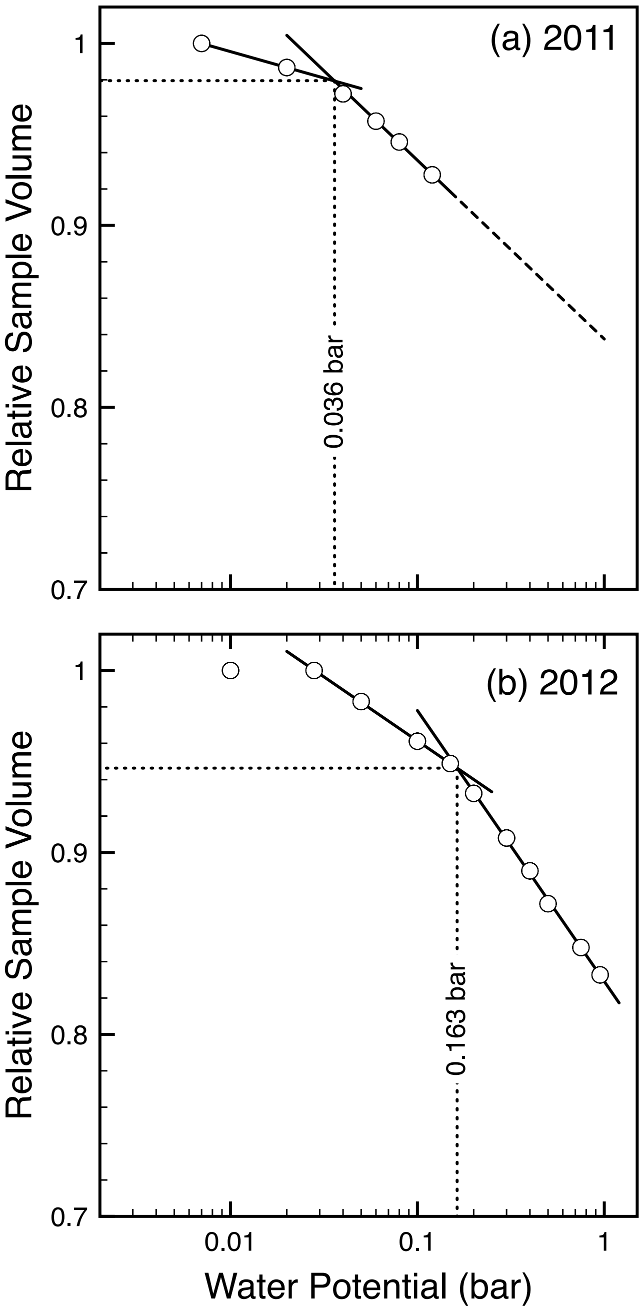

Figure 5-4 Consolidation test of 2011 and 2012 soil cores. Historic limit of dryness in both samples is illustrated by the dotted line.

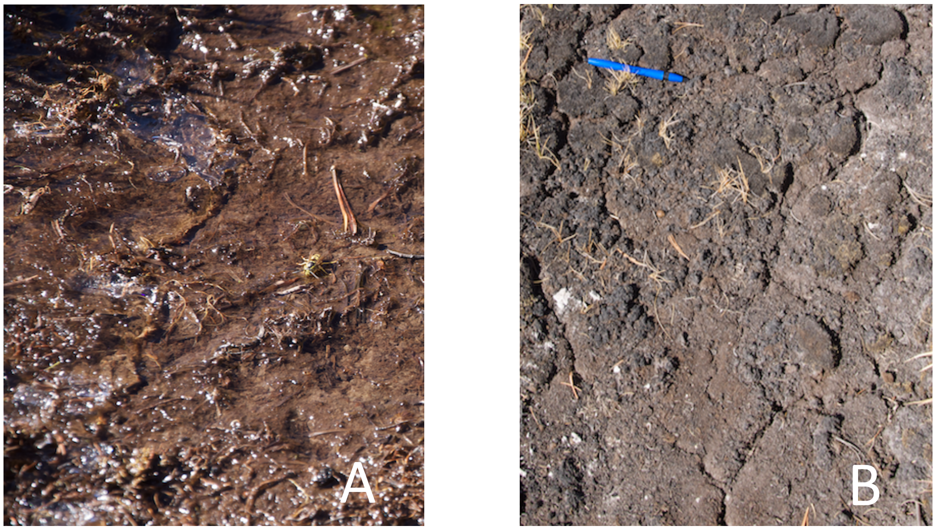

Figure 5-5 Pictures of meadow peat soil from study site at saturation (a) and at the end of the summer in 2012 (b).

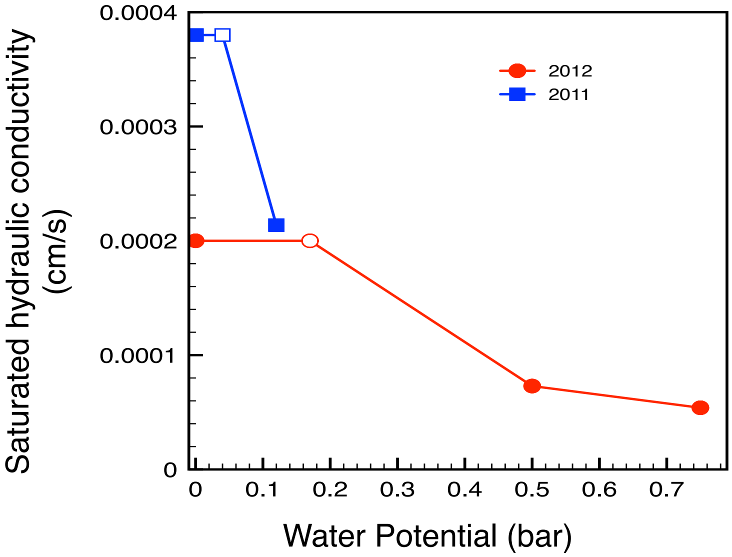

Figure 5-6 Change in saturated hydraulic conductivity in 2011 and 2012 peat samples. The open symbols represent the saturated hydraulic conductivity at the respective historic limits of dryness for 2011 and 2012.

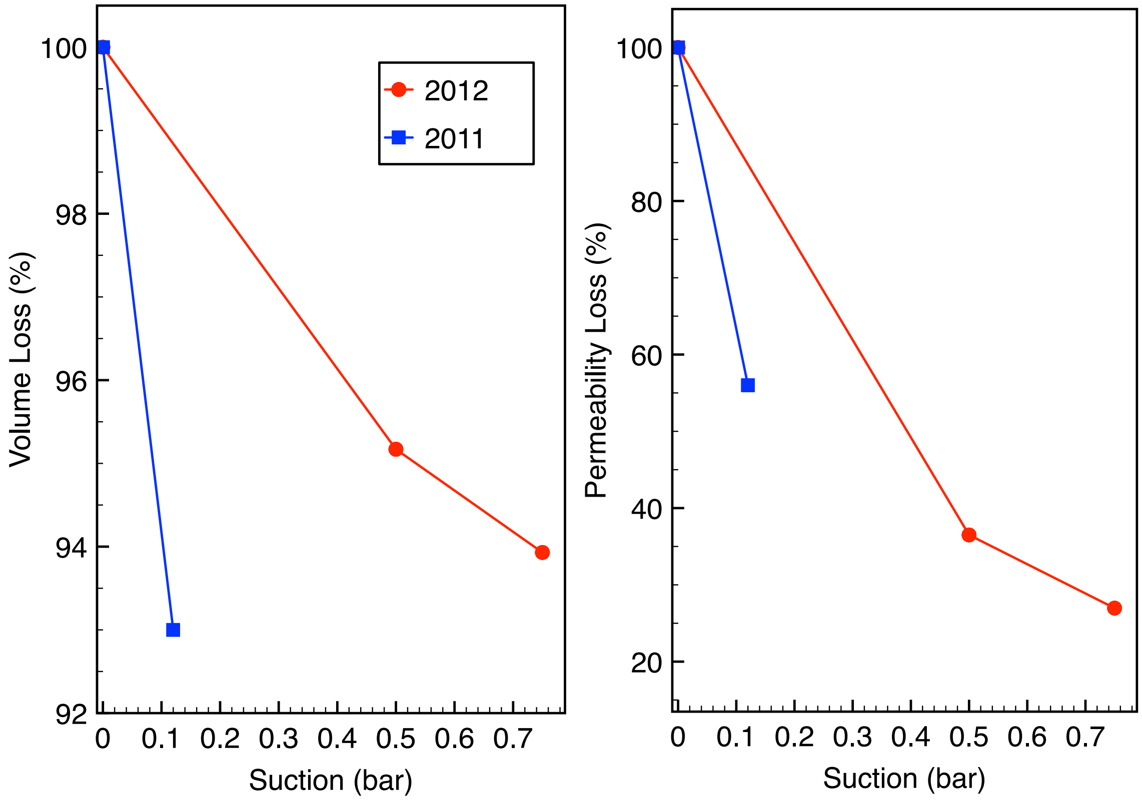

Figure 5-7 Porosity and permeability loss as a result of consolidation in 2011 and 2012.

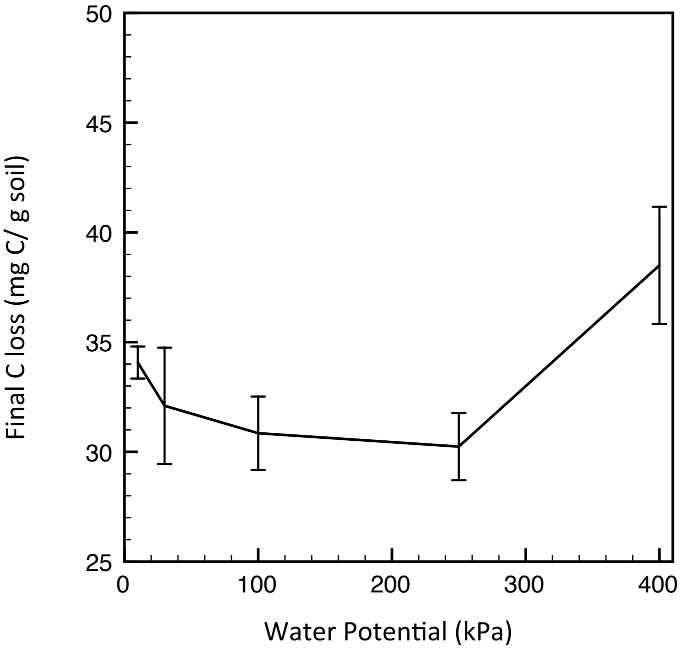

Figure 5-8 Total carbon loss from long-term incubation of the composited peat soil (0-10 cm). The point is the total carbon loss from the intact cores (0-5 cm). Error bars represent the standard error of the three replicates.

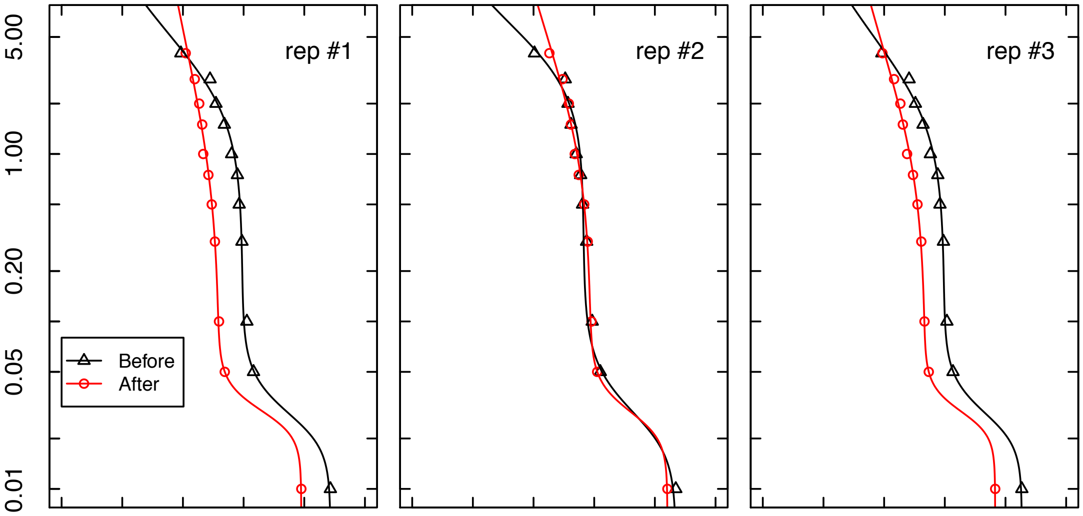

Figure 5-9 Before and after the incubation, intact paired peat cores from 2011 were used to generate water retention curves fitted with the Durner-vanGenuchten model for a bimodal pore distribution. The three replicate pairs are plotted independently.

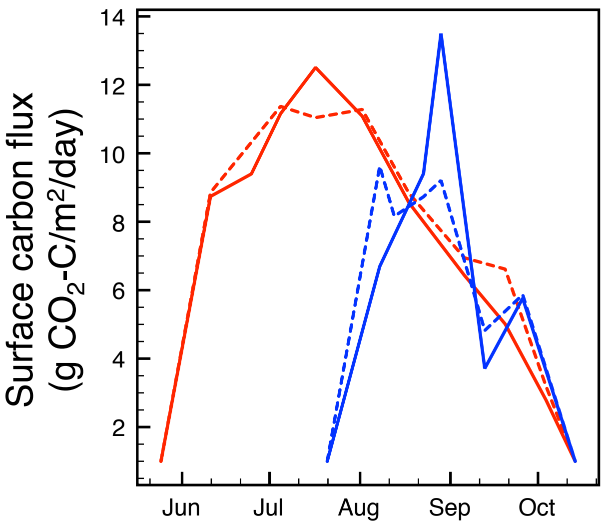

Figure 5-10 Surface CO$_2$ flux from the peat soil study site in 2011 and 2012. Dashed and solid lines represent two separate study sites (Hall A-solid and Hall B-dashed) that were sampled. Measurements began as soon as the meadow was snow-free each year.

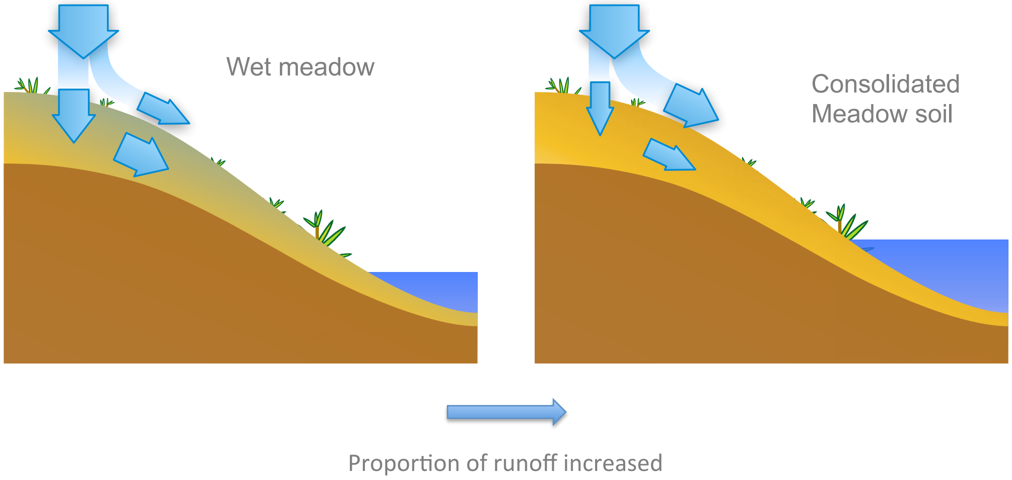

Figure 5-11 Conceptual diagram of the impacts of consolidation on ecosystem services.

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CHAPTER 6. Conclusion

High elevation meadows are a keystone feature in mountain regions through their role in the storage and movement of snowmelt to the river systems. Crucial to this ability is the amount of carbon stored in the meadow soils that facilitates the retention of water and contributes to the physical structure of the meadow soils. While meadow systems are fairly stable entities in a mountain landscape, how they will respond to future climate change is uncertain and can have large implications for downstream communities.

This dissertation focused on increasing understanding of how changes in seasonality (ie. an early versus a late spring), and the timing and duration of snowpack can influence resilience of meadow soils. In Chapter 2 we found that an early spring increases the length of growing season in the meadow, but can damage meadow vegetation through severe frost events. The loss of productivity coupled to drydown over the long summer season contributed to an increase in soil respiration that shifted the meadow from a sink to a source of carbon. By analyzing several years of data, it is apparent that environmental parameters such as the timing of "snow-on" and "snow-off" in the meadow can influence ecosystem processes such as primary productivity and soil respiration in the meadow, in that what happens during the winter months will be relevant the following summer season. Years where there is no continuous snowpack for much of the winter can significantly impact the ecosystem processes such as primary productivity and soil respiration.

With a shift to an earlier spring, and little summer precipitation, high elevation meadows are faced with the potential for increased drying. To better understand the implications of this drying on decomposition of soil organic matter in the meadow, we incubated soils from across the hydrologic gradient at five different water potentials (Chapter 3). We found the largest cumulative carbon mineralization at the highest (wettest) and lowest (driest) water potential. This was counter to what was expected at the lowest water potential and led us to develop a conceptual model of decomposition in these types of organic soft soil. In this model, after the large pores are drained at the highest water potential, the newly aerated conditions enhance decomposition of organic matter. However, if drying proceeds, the soil matrix begins to consolidate and rather than allowing continual drainage of water, there is a redistribution of water in the soil matrix as meso/micro pores increase at the expense of the macropores. This redistribution of water keeps the smaller pores saturated at lower water potentials, which reduces the rate of decomposition. As the tensile strength of the soil increases, the soil matrix stiffens and no longer consolidates. At this point, the remaining water begins to drain and there is a subsequent peak in carbon mineralization. Although this was a laboratory-based incubation, it does better inform us as to the expected response of the ecosystem when exposed to drying in the field. While some of the soils aren't significant, they all show a clear trend to increased flux at the dry end. There is clearly a mechanism occurring that hasn't previously been investigated. The importance of such a threshold could prove critical in understanding how organic soils will respond to future drying.

In order to quantify the reduction in volume of these soils that occurs with desiccation, we developed a method to induce soil consolidation through capillary suction (Chapter 4). This method is important in that it allows for simultaneous desaturation of a soil sample and measurement of soil volume change. To the best of our knowledge, this is the first of this kind of method that directly allows for determination of the historic limit of dryness in a soil. By draining the soil by capillary suction we are better able to mimic the desiccation that would occur in the field. This allows for prediction of how soils will respond to future changes in climate.

We tested this method of consolidation by capillary suction on intact cores collected in 2011 and 2012 from a peat meadow in the Hall RNA. We found that the historic limit of dryness had shifted from 0.036 to 0.163 bar from 2011 to 2012. With this change in the historic limit of dryness was a loss of soil volume and permeability as well as predictions for losses with future desiccation of the meadow soil.

This research illustrates how sensitive high elevation meadow soils are to a change in seasonality. The timing and duration of snowpack and subsequent snowmelt drives the onset of the growing season in the meadow. Changes to the timing of the growing season can cause a meadow system to shift from being a sink to a source of carbon. In addition, if the meadow desiccates due to drying over a prolonged growing season, or lack of snow cover in the winter, structural changes in the soil can cause rapid and irreversible loss of porosity and permeability. This translates to less infiltration into the soil during the spring snowmelt period and more surface runoff to the streams. These findings also highlight the need for future research and monitoring of high elevation meadow soils to determine the long-term implications for the soil structural changes as well as significant carbon losses over the past several years.